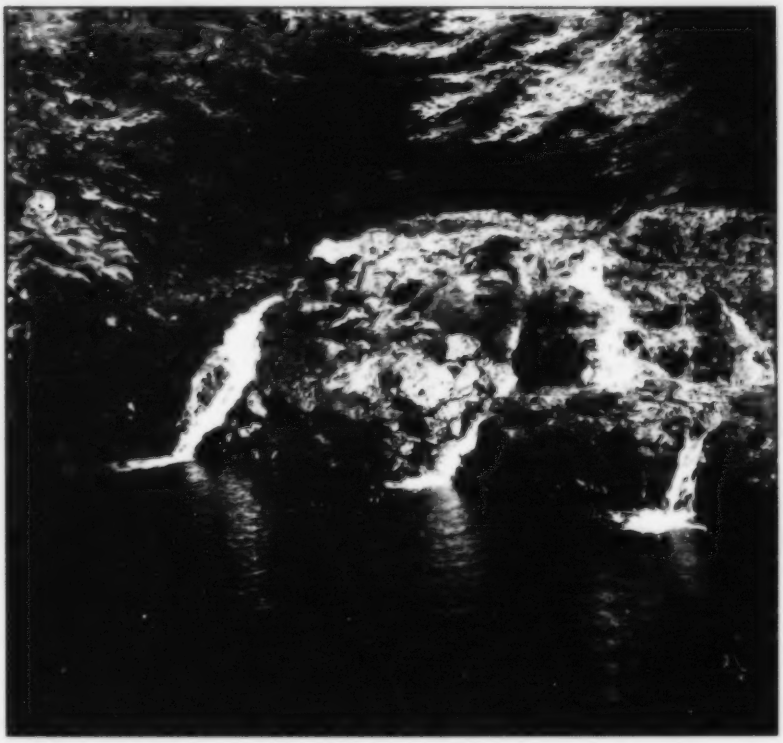
MRR 2006–3

New Nouveau Brunswick

Natural Resources
Minerals, Policy and Planning

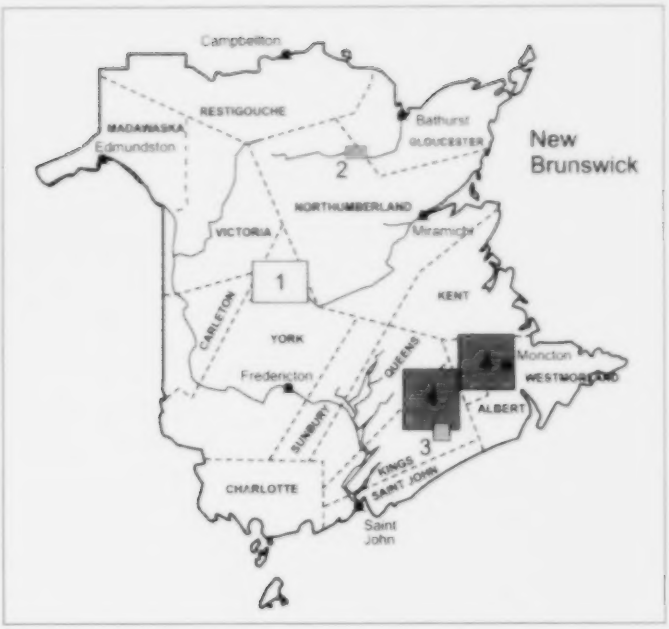
Mineral Resource Report

GEOLOGICAL INVESTIGATIONS IN NEW BRUNSWICK FOR 2005



Gwen L. Martin, editor

2006



Mineral Resource Report 2006-3

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Numbers on location map refer to papers in this volume, as listed in *Table of Contents* (facing page).

Cover illustration *Waterfalls over indurated and fractured mudstones and sandstones of the McQuade Brook Formation, McQuade Brook, southeastern New Brunswick.*

Photograph courtesy of Clinton J. St. Peter

Figure preparation Terry Leonard

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Hon. Donald Arseneault
Minister of Natural Resources

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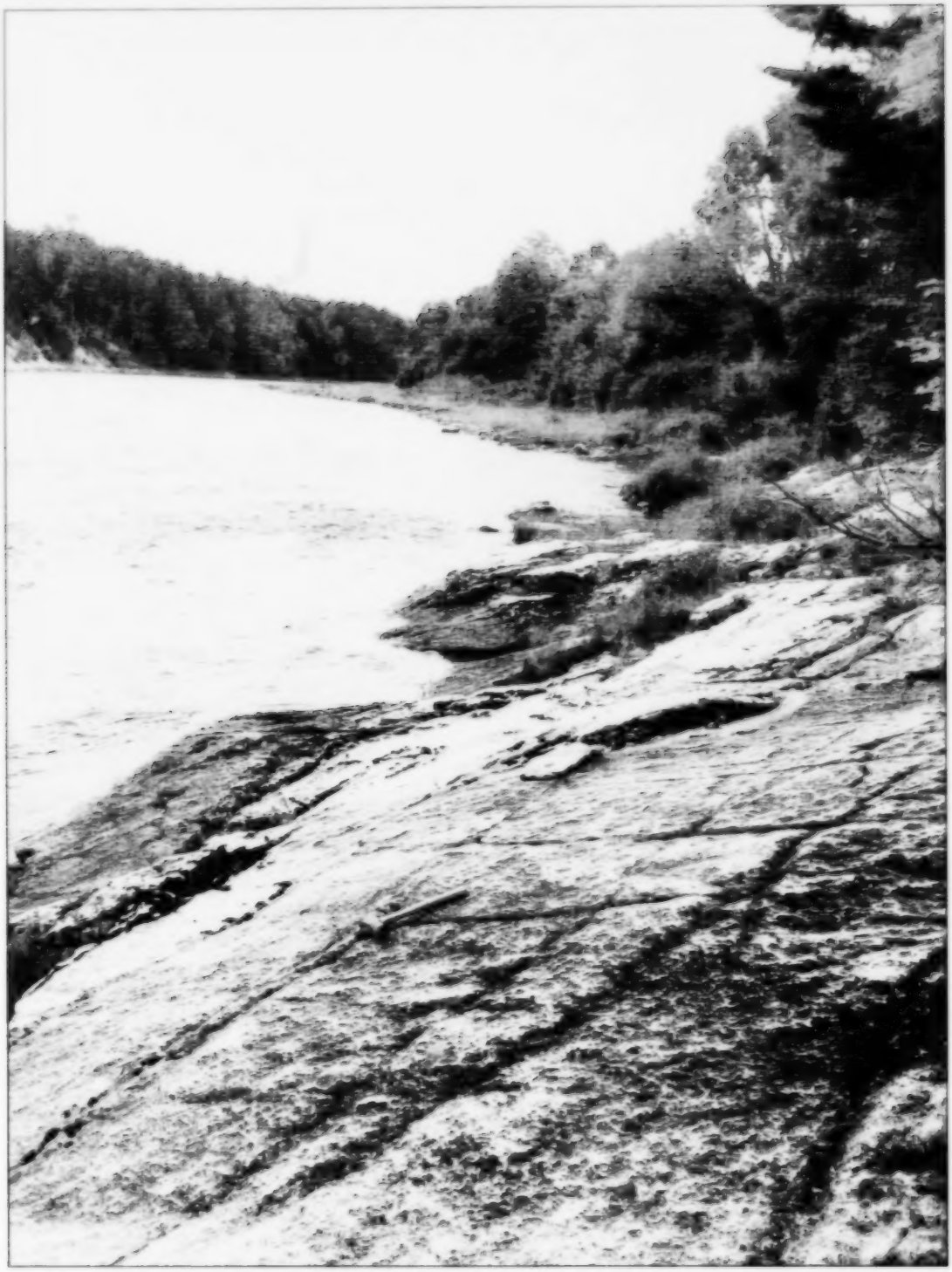

Frontispiece. Glaciated outcrop exposes the Trout Lake lobe of the Dungarvon Granite along the Southwest Miramichi River, central New Brunswick. Outcrop occurs in the map area covered by the following report on the Gaspereau Ice Centre in New Brunswick (see facing page). *Photograph courtesy of Kay G. Thorne*.

A New Interpretation of the Late Glacial History of Central New Brunswick: The Gaspereau Ice Centre as a Younger Dryas Ice Cap

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Seaman, A.A. 2006. A new interpretation of the late glacial history of central New Brunswick: the Gaspereau Ice Centre as a Younger Dryas ice cap. *In* Geological Investigations in New Brunswick for 2005. *Edited by* G.L. Martin. New Brunswick Department of Natural Resources; Minerals, Policy and Planning Division, Mineral Resource Report 2006-3, p. 1–36.

A till formed during the Younger Dryas Chronozone was identified by the Geological Survey of Canada in the 1980s at a site near Todd Mountain in the Hayesville (NTS 21 J/10) map area of central New Brunswick. The Younger Dryas was a millennial-scale cool interval that interrupted the general warming trend at the very end of the Pleistocene. New data, mainly from the Hayesville map area, suggest that the till was formed by a glacier centred to the southeast over the New Brunswick Lowlands. This glaciation is attributed to the Gaspereau Ice Centre that previously was ascribed to the main Late Wisconsinan glaciation, but now is believed to have been active as an identifiable ice centre primarily during the Younger Dryas.

La Commission géologique du Canada avait relevé, dans les années 80, un till s'étant formé au cours de la chronozone du Dryas récent en un emplacement situé près du mont Todd, à l'intérieur du secteur cartographique de Hayesville (SNRC 21 J/10) dans le centre du Nouveau-Brunswick. Le Dryas récent a constitué à l'échelle du millénaire un intervalle frais qui a interrompu la tendance générale au réchauffement vers la toute fin de l'époque pléistocène. De nouvelles données, provenant principalement du secteur cartographique de Hayesville, laissent supposer que ce till a été formé par un glacier centré dans le Sud-Est au-dessus des basses terres du Nouveau-Brunswick. Cette glaciation est attribuée au centre glaciaire de Gaspereau, qui avait antérieurement été rattaché à la principale glaciation du Wisconsinien tardif, mais qui, d'après ce que l'on croit maintenant, aurait été un centre glaciaire identifiable actif principalement au cours du Dryas récent.

Introduction

In their summary of the Quaternary geology of New Brunswick, Rampton et al. (1984) indicated that central New Brunswick (Fig. 1) was completely glaciated at the maximum of the Bantalor phase, about 13 400 BP, largely by ice of the Gaspereau Ice Centre. (Note: all

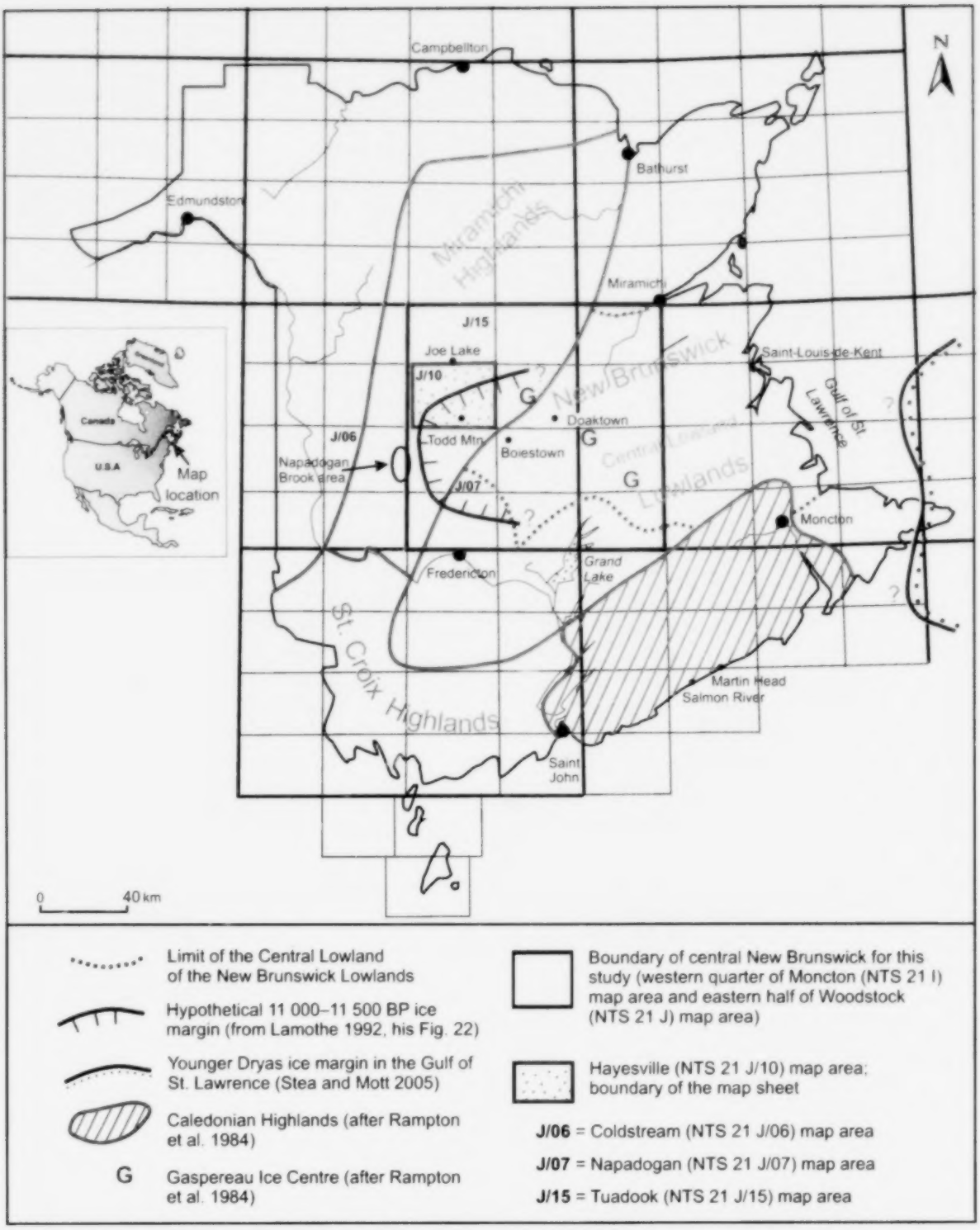
Campbellton
Bathurst
Edmundston
Miramichi Highlands
Miramichi
J/15
Joe Lake
J/10
Saint-Louis-de-Kent
New Brunswick
Central Lowlands
Gulf of St. Lawrence
Doaktown
Todd Mtn
Boiestown
J/06
Napadogan Brook area
J/07
Moncton
Fredericton
Grand Lake
St. Croix Highlands
Martin Head
Salmon River
Saint John
Canada
U.S.A
Map location
N
0
40 km
Limit of the Central Lowland of the New Brunswick Lowlands
Hypothetical 11 000–11 500 BP ice margin (from Lamothe 1992, his Fig. 22)
Younger Dryas ice margin in the Gulf of St. Lawrence (Stea and Mott 2005)
Caledonian Highlands (after Rampton et al. 1984)
G Gaspereau Ice Centre (after Rampton et al. 1984)
Boundary of central New Brunswick for this study (western quarter of Moncton (NTS 21 I) map area and eastern half of Woodstock (NTS 21 J) map area)
Hayesville (NTS 21 J/10) map area; boundary of the map sheet
J/06 = Coldstream (NTS 21 J/06) map area
J/07 = Napadogan (NTS 21 J/07) map area
J/15 = Tuadook (NTS 21 J/15) map area

Figure 1. Location map outlining the New Brunswick physiographic regions (after Rampton et al. 1984) and National Topographic System (NTS) map areas referred to in the text. The map also delineates the Gaspereau Ice Centre, and margins of the proposed Younger Dryas ice caps.

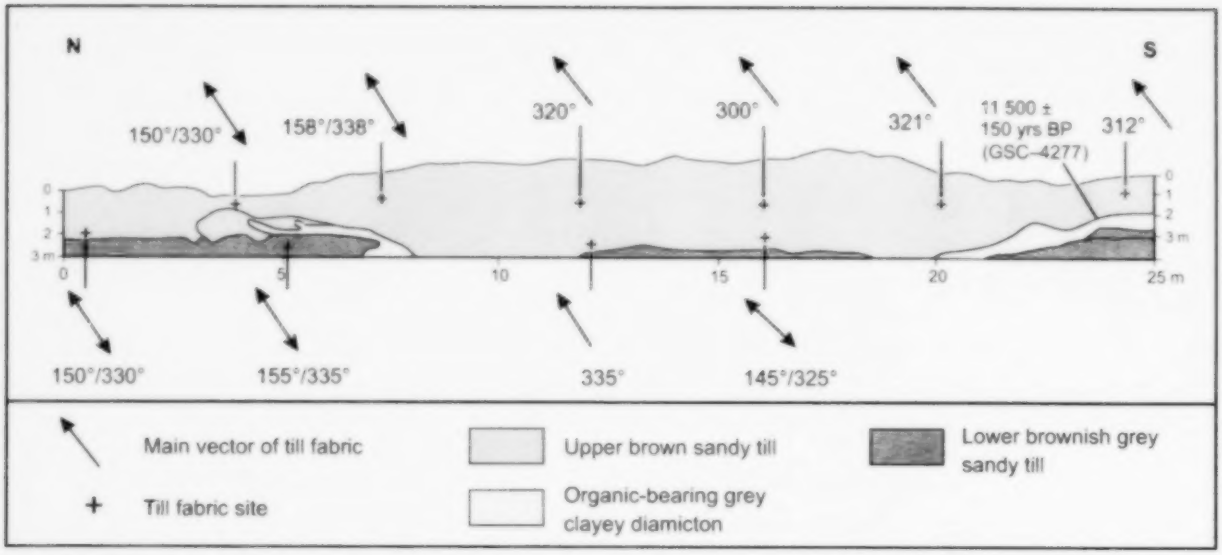

Figure 2. Stratigraphy at the Todd Mountain organic site, GSC trench 2009 (after Sacré 1989 and Lamothe 1992).

cited dates are in radiocarbon years before present.) The Gaspereau Ice Centre was one of several ice centres believed to have interacted during the Late Wisconsinan to form the complex record of glaciation in New Brunswick (Rampton et al. 1984, their Fig. 31). By the subsequent Millville/Dungarvon phase, about 12 700 BP, the North Maine Ice Divide in the northwestern and western parts of the province was the only glacier interpreted to remain in New Brunswick (Rampton et al. 1984, their Fig. 34). Rampton et al. (1984) indicated that the available data suggested that the province was completely deglaciated by 12 000 BP. In contrast, contemporary dates for the deglaciation of adjacent areas were given as before 11 200 BP for northwestern Maine (Lowell 1985); by 10 500 BP for northeastern Maine (Newman et al. 1985); around 9800 BP for the Gaspé Peninsula (David and LeBuis 1985); and into the Holocene (after 10 000 BP) for Nova Scotia (Stea 1984).

During the late 1980s, a Geological Survey of Canada (GSC) till trenching/geochemistry project conducted in the Hayesville (NTS 21 J/10) map area of central New Brunswick (Fig. 1) found evidence that glaciers had persisted in the Miramichi Highlands well after 12 000 BP. Organic sediments dating at 11 500 BP were encountered below 2 m of surface till in GSC trench 2009 (Fig. 2) near Todd Mountain (Sacré 1989; McNeely and McCuaig 1991; Lamothe 1992). Lamothe (1992) recognized this was "a critical site for the interpretation of late glacial climatic changes in Eastern Canada" (Lamothe 1992, p. 30), and he inferred that reactivation of the ice that formed the surface till at this site was correlative with Younger Dryas climatic deterioration.

In Lamothe's (1992) interpretation of the depositional context at GSC trench 2009, he presented two alternate scenarios, referred to here as interpretation 1 and interpretation 2. He noted that the upper till contains Devonian granitic clasts from outcrop to the west in the Central Miramichi Highlands, but that most till fabrics at the site are oriented toward the northwest. He stated: "The interpretation is that easterly moving ice deposited the upper till

because its lithological content reflects a western provenance....Upon melting, the structures in the till were reoriented by movement down the regional topographic slope...." (Lamothe 1992, p. 26, interpretation 1). On the other hand, he also stated: "The eastern flank of the Miramichi Highlands and, presumably, parts of the New Brunswick Lowlands were, however, occupied by a mass of residual ice that was still active, as evidenced by the buried organic material at Hayesville...." (Lamothe 1992, p. 26, interpretation 2), which would imply a northwestward ice flow in the Todd Mountain area. Moreover, Lamothe (1992, his Fig. 22) illustrated a hypothetical 11 500–11 000 BP western margin for such an ice mass (as shown here on Fig. 1).

The residual ice mass location plotted by Lamothe (1992) approximates that given by Rampton et al. (1984) for the Gaspereau Ice Centre. Subsequent research has resulted in a different glaciation model for southern and central New Brunswick, one that involves the migrating outflow centre of a single regional ice mass, rather than multiple interacting centres (Seaman et al. 1993; Seaman 2004). However, as noted by Seaman (2004, p. 163), "Local ice caps probably formed during the waning stages of the main Late Wisconsinan glaciation, and may have reformed or enlarged during the Younger Dryas." The Gaspereau Ice Centre may be one of these Younger Dryas ice caps.

This paper summarizes the evidence for Younger Dryas glaciation in southern and central New Brunswick and then presents new evidence, primarily from the Hayesville map area (Fig. 1), that supports the presence of an active Gaspereau Ice Centre during the Younger Dryas Chronozone.

The Younger Dryas in the Maritime Region

The Younger Dryas Chronozone represents a time of climatic cooling at the end of the Pleistocene (about 11 000–10 000 BP) and is long known from northwestern Europe and the North Atlantic Ocean (Mayle and Cwynar 1995a). It was the last of several climatic oscillations that characterized the end of the Late Wisconsinan glaciation (Dansgaard et al. 1989). It followed the Allerød interstadial — a warm period preceded by the Older Dryas stadial, or cool period (Clapperton 1995) — and was followed by the pre-Boreal (Dansgaard et al. 1989), the first part of the current Holocene interglacial. The name 'Younger Dryas' was derived from the recognition (in Scandinavia) of fruits and leaves of the tundra plant *Dryas octopetala* in buried organic sequences in which the plant was stratigraphically overlying evidence for vegetation by shrubs and trees during the Allerød (Peteet 1995).

The first convincing evidence for the Younger Dryas event in the Maritimes region was presented by Mott et al. (1986); the evidence was based on studies of lake sediments from a few New Brunswick sites and of both lake sediments and buried organic sites from Nova Scotia. The authors noted that, at some Nova Scotia sites, "overlying till-like diamictons

suggest the possibility of glacial activity, but reactivation or reglaciation has not been demonstrated..." (Mott et al. 1986, p. 248). Further studies at Nova Scotia sites have since confirmed the presence of Younger Dryas glaciers over highland areas (Stea and Mott 1989, 1998, 2005; Mott and Stea 1993; Stea 2004). This glacial episode has been called the Collins Pond phase after a key site in Nova Scotia (Stea 2004).

Stea (2004) and Stea and Mott (2005) also presented evidence that a significant Younger Dryas ice cap was centred over Prince Edward Island. This glacier flowed southward into northern Nova Scotia, forming a fluted till sheet 2 m to 10 m thick. According to Stea (2004, his Fig. 9), the southern margin of this ice cap lay along the northern margin of the Cobequid Highlands. The westward extent of the ice sheet into New Brunswick is unknown, but Stea and Mott (2005, their Fig. 12) indicated that it covered at least the eastern part of the Tormentine Peninsula (Fig. 1). Rampton et al. (1984; their Fig. 27) identified a southerly trending "late flow of decaying ice caps" along the northern shore of the Tormentine Peninsula that may relate to this Younger Dryas ice flow.

In Maine, the Late Wisconsinan glacier disintegrated into two large remnants occupying lowland areas in the northern part of the state. These two ice masses either stabilized or readvanced during the Younger Dryas (Borns et al. 2004). In contrast, the Notre Dame Mountains of the Gaspé Peninsula remained under continuous ice cover until the end of the Younger Dryas, undergoing final deglaciation in the early Holocene (Richard et al. 1997). In a recent study of sediment core from Island Lake in northern New Brunswick, Mott et al. (2004) reported an absence of evidence for the Younger Dryas vegetation change and suggested that organic sedimentation in the lake began just before 10 000 BP. This age corresponds well with the interpretation of Richard et al. (1997).

Given the existence of Younger Dryas-age ice caps over both highland and lowland areas, and on all sides of New Brunswick except to the southwest, glaciers of similar age almost certainly must have been present within the province.

THE YOUNGER DRYAS IN NEW BRUNSWICK

Southern New Brunswick

Mott (1975) first noted a late glacial climatic deterioration in New Brunswick, basing his observation on pollen sequences in cores from Basswood Road Lake (now called Splan Pond; see Seaman et al. 1993) and Little Lake in southwestern New Brunswick. Several subsequent studies of lake sediment profiles from New Brunswick sites (Rawlence 1988; Rawlence and Senior 1988; Levesque et al. 1993a, 1993b, 1994, 1996, 1997; Mayle et al. 1993a, 1993b; Cwynar et al. 1994; Mayle and Cwynar 1995a, 1995b) have clearly documented the presence of a cool interval correlative with the Younger Dryas. Most of the

studied sites are in southwestern New Brunswick, but one site (see *Joe Lake*, below) occurs at Joe Lake in central New Brunswick, 22 km north of the Todd Mountain site (GSC trench 2009) mentioned above and described below. These studies are particularly significant, because the lake basins lack evidence of occupation by a Younger Dryas glacier, thereby putting constraints on the extent of such a glacier.

Further constraints on the extent of any Younger Dryas glacier are imposed by the existence of preserved, intricate esker systems in southwestern New Brunswick (Seaman 2004). The systems feed into the Pennfield–Utopia and Pocologan delta complexes east of St. George (Fig. 3; Seaman et al. 1993). A date of 13 000 BP on shells from the toe of the younger part of the Pennfield–Utopia delta complex (Gadd 1973; Seaman et al. 1993) indicates that this delta complex and its feeder esker system were formed during the Bantalor phase of Rampton et al. (1984), well before the Younger Dryas. Therefore, any incursion of Younger Dryas glaciers into southwestern or west-central New Brunswick would have been restricted to areas not occupied by this relict esker system.

In the Saint John area (Fig. 1), Nicks (1991) identified a possible glacial readvance after 11 300 BP on the basis of marine shells dated at 11 620 ± 250 BP (BGS 1223) in an overturned, black, stratified, fine sand layer overlain by a thin (0.30–0.50 m) reddish brown diamicton, which in turn is overlain by gravelly, glaciofluvial foreset beds. She suggested that ice advanced southwestward down the Hammond River valley east of Saint John, filling Saint John Harbour and impinging on the southern side of the Sheldon Point Moraine.

Foisy and Prichonnet (1991) suggested that a residual ice cap overlay the Caledonian Highlands at the end of the Late Wisconsinan (Fig. 1), and inferred that it could be related to the Younger Dryas glacigenic deposits reported from Nova Scotia by Stea and Mott (1989). The residual ice cap could have expanded and spread northward onto the adjacent New Brunswick Lowlands to become the source of the glacier inferred by Nicks (1991); moreover, the expanded ice cap could have dammed the Saint John River valley to form Glacial Lake Acadia, as has been postulated by Seaman (2004).

Further possible evidence for a Younger Dryas glacier over the Caledonian Highlands was found in a Bay of Fundy coastal section just east of Salmon River (Fig. 1). P. Finamore (field notes, 1976) reported 6.1 m of cobbly sandy gravel overlying 1.2 m of contorted, stratified silt. Seaman (1988, site H6-101) used Finamore's description to suggest that the deposit was a remnant of a glaciofluvial delta and that the section could be part of a slump block, thereby accounting for contortion in the silt. When the author visited the site in 1988, the base of the gravel was no longer exposed. However, it was clearly underlain, at an unknown depth, by reddish glaciomarine clay characteristic of parts of the Bay of Fundy area that were submerged during the late glacial sea-level highstand. Barnacle shells from similar clay at Martin Head to the east have been dated at 13 340 ± 400 BP (BGS 1407;

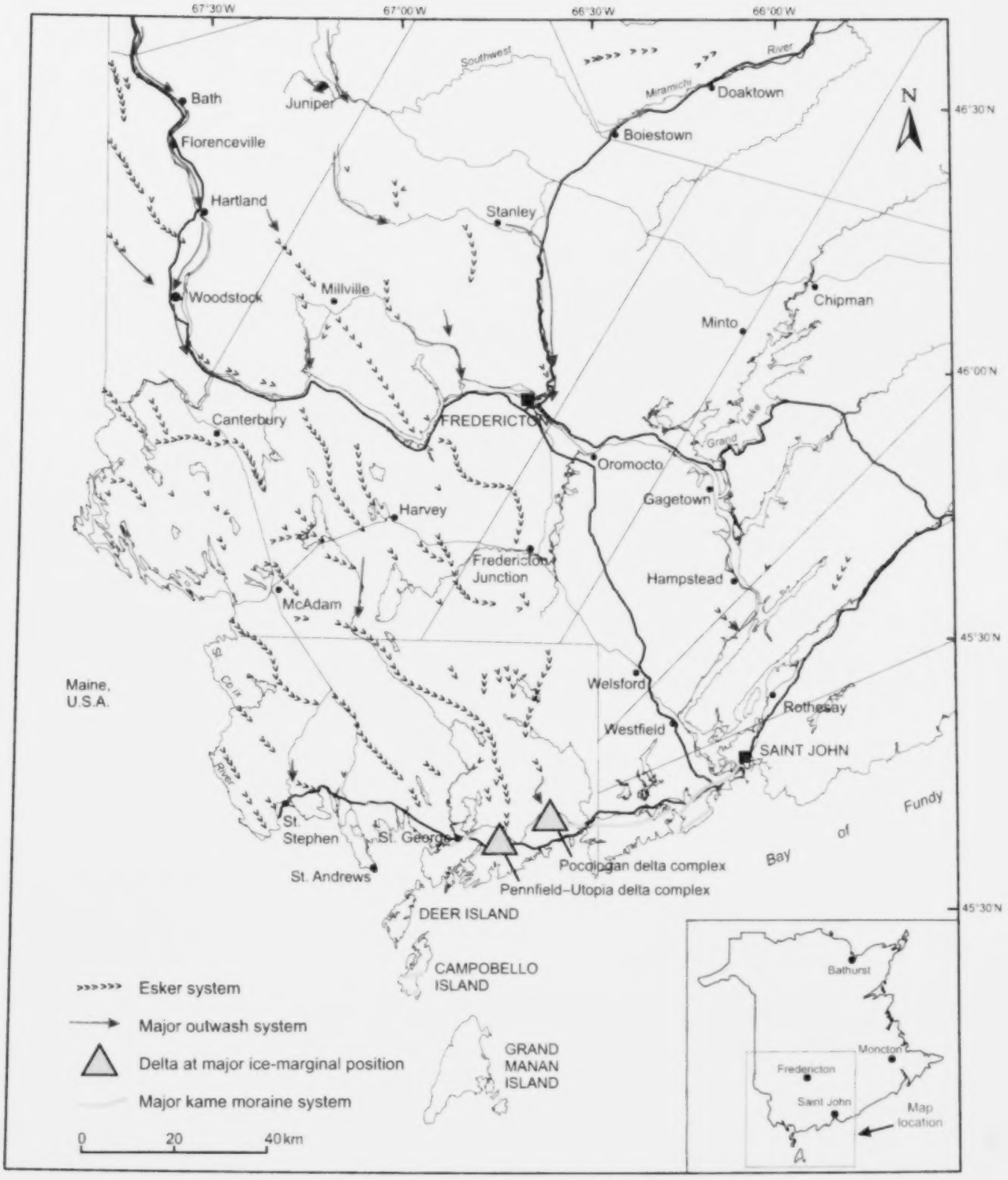

Figure 3. Meltwater drainage pattern in southwestern and west-central New Brunswick (from Seaman et al. 1993).

uncorrected age), and similar dates for glaciomarine deposits have been reported from the Saint John area (see Nicks 1991 and Seaman et al. 1993).

The 1988 Salmon River section just mentioned exposed a cut through the fill in a small kettle hole in glaciofluvial gravel (Fig. 4). The kettle fill comprised primarily organic silt and clay, with large pieces of wood in the lower part of the unit. If this glaciofluvial gravel had been related to an earlier Late Wisconsinan glacial phase, the expected date for organic material near the kettle-hole base would be in the order of 12 000 BP. Instead, one piece of wood gave a date of 8730 ± 120 BP (BGS 1259; uncorrected age), which strongly suggests that the glaciofluvial delta is of Younger Dryas age. If so, the contortion noted by Finamore could be due to glacial overriding during the Younger Dryas.

Figure 4. The Salmon River kettle-fill section. From top to bottom, the section consists of 0.2 m of peaty soil; 0.3 m of organic clay; 0.1 m of poorly sorted gritty sand; 0.3 m of organic silt and clay containing large pieces of wood (8730 ± 120 BP; BGS 1259); and 1.0+ m of bouldery, cobbly gravel. The scale card is 85 mm long.

Central New Brunswick

Todd Mountain Area

The Todd Mountain site (GSC trench 2009) lies south of the Southwest Miramichi River at the northeast end of the Todd Mountain ridge, approximately 2.5 km northeast of the peak and about 23 km west-northwest of Boiestown (Fig. 1). With 2+ m of till overlying organic material dated at 11 500 BP, the site provides strong evidence for an active Younger Dryas glacier in central New Brunswick (Fig. 2). McNeely and McCuaig (1991) and Lamothe (1992) reported the site location as 46°33'N, 66°41'W. More precise coordinates (46°33.4'N Lat., 66°40.7'W Long.) were determined by transposing the site location on Plate 1 of Lamothe (1990) onto the Hayesville (NTS 21 J/10) 1:50 000 scale sheet.

The organic material, described as a black silty peat, was enclosed in fine-grained clayey diamicton beneath 2 m of an upper sandy till at the south end of a trench 25 m long (Fig. 2; Sacré 1989; Lamothe 1992). Elsewhere in the trench, this upper till is greater than 5 m thick (Fig. 2). The organic-bearing diamicton overlies a lower sandy till that Lamothe (1992) correlated with the older Late Wisconsinan till, which occurs at surface to the southwest. In contrast, Sacré (1989) believed that all till exposed in the trench was a single till, and that the organic-bearing diamicton was transported from elsewhere and deposited within this till.

Till-fabric trends were measured at ten locations in GSC trench 2009: nine in the 'upper till' and one in the 'lower till' (Fig. 2). Sacré (1989) reported northwest trends for the fabric at five locations in the upper till, including the two locations directly overlying the organic-bearing material. The remaining four fabrics in the upper till and the fabric in the lower till displayed trends that could be either northwest or southeast. At the fabric sites, the pebble component of the upper till varied from 4% to 13% granite, whereas the pebble component of the lower till was 7% granite (Lamothe 1992). The granitic pebbles consisted of clasts from Devonian granites that outcrop to the north and west. This granite content led Lamothe (1992, interpretation 1) to suggest that the upper till was deposited by eastward-moving ice. He also noted that "the presence of a shear structure in the northern end of the trench, the occurrence of clay skins around the pebbles of the till, and its high primary porosity, all tend to suggest the glacial debris was released by basal meltout. Upon melting, the structures in the till were reoriented by movement down the regional topographic slope...." (Lamothe 1992, p. 26).

Although Lamothe (1992) interpreted the shear structure at the north end of GSC trench 2009 (Fig. 2) as indicative of basal meltout, Fyffe and Pronk (1985, their Fig. 29) interpreted a similar feature in the Tuadook Lake (NTS 21 J/15) map area to the north as a glaciotectonic structure. They attributed its formation to glacial shearing, implying active ice flow under conditions of water saturation at the base of the glacier. Glacigenic shearing or deformation by active ice was also documented at other New Brunswick locations (Broster and Seaman 1991; Broster and Park 1993). In addition, Sacré (1989, p. 39) noted that the lenses of organic material in GSC trench 2009 sampled for dating "are thin and discontinuous, and appear to have been sheared in the grey, clayey diamicton." Therefore, subglacial deformation by an active, northwest-flowing glacier would seem to be a better interpretation of the shear structure at the trench's north end.

Test Pit HV-P11.5

To gain a better understanding of stratigraphy in the Todd Mountain area, test pit HV-P11.5 was excavated in a new roadcut about 150 m north of GSC trench 2009 (Seaman et al. 2005). The pit was on the northwest-facing slope of a northeast-trending valley, and the exposed section comprised 0.6+ m of rubbly deformation till overlain by 1.2 m of sandy basal till (Fig. 5). The basal till varied from moderately compact to compact near the base of

Figure 5. A till section (site HV-P11.5) situated 150 m north of the GSC trench 2009 near Todd Mountain. The head of the rock pick rests on the top of the deformation till; the base of the marker pen rests on the top of the 'mixing zone' between the upper basal till and the deformation till. The rock hammer is 33 cm long.

the soil BC-horizon to compact to very compact in the 'mixing zone' in the basal 0.2 m, where material from the underlying deformation till was incorporated. There was no evidence of the lower sandy till identified in GSC trench 2009, which suggests that Sacré (1989) may have been correct in her interpretation that only one till was present.

Deformation till at the test pit consisted of 98% angular metasedimentary clasts of local derivation and contained no granite clasts. The mixing zone at the base of the sandy basal till contained 87% angular local clasts and 1% granite, indicating reworking (or mixing) of material from the deformation till into the base of the sandy basal till. The upper sample from the overlying sandy basal till contained 9% Devonian granitic clasts and so is comparable to till in GSC trench 2009.

The till fabric in the mixing zone (site HV-P11.5-2 in Table 1) is comparable to fabrics reported by Sacré (1989) for GSC trench 2009 and suggests deposition by northwestward-flowing ice. The till fabric in the upper part of the till (site HV-P11.5-1 in Table 1) is parallel but indicates a reverse trend toward the southeast. An explanation for this can be derived from examination of the depositional context of the till. The local slope is approximately 12° toward the northwest, as indicated by the 1:50 000-scale NTS topographic map. Because till-fabric plunges are measured with respect to the horizontal, the plunge of the fabric in the upper part of the till (12°) is essentially parallel to the local slope. In contrast, fabric in the mixing zone has a strong southeast (upslope) plunge of about 17° relative to the local slope. The interpretation offered here is that the mixing zone fabric represents basal lodgement by a northwestward-flowing glacier, while the upper fabric retains the approximate orientation of the ice flow, but has had its clasts rotated into a slope-parallel alignment by basal meltout or by post-glacial downslope movement.

As noted above, Lamothe (1992, interpretation 1) inferred eastward ice flow for the Younger Dryas till on the basis of granitic clasts; he also attributed the northwest trend of the till fabrics to later downslope movement. However, an alternate explanation is that the till

Table 1. Till fabric trends in central New Brunswick.

Grid Point No.	Site No.	Easting NAD27	Northing NAD27	n*	Mean up-dip a-axis trend	Std. Dev.	Mean angle of imbri-cation†	Topographic Slope‡ Exposure	Slope angle
CS-T04.5	AS030025	648050	5144230	30 (3)	262°	23°	33°	SW	4°
CS-T09-2		648020	5134750	50 (9)	142°	28°	4°	crest	0°
CS-TV03		649270	5147090	30 (5)	333°	32°	15°	NW	5°
CS-V08-2		650050	5137190	30 (7)	035°	19°	19°	NE	11°
CS-V10.5a-1		649640	5132320	50 (2)	265°	47°	14°	SW	8°
CS-V10.5a-3		649640	5132320	50 (4)	261°	33°	2°	SW	8°
CS-VW02-2	AS030013b	651020	5149480	50 (4)	322°	29°	5°	NNE	3°
CS-W07		651850	5139050	50 (6)	192°	21°	23°	WNW	6°
CS-W08.5		651350	5136020	50	095°	33°	28°	E	6°
CS-W09		651950	5134960	60 (6)	267°	38°	33°	crest	0°
CS-W10b		651560	5133220	50 (1)	043°	31°	14°	NE	7°
CS-WX03a	AS030011	653150	5146890	50 (7)	149°	32°	31°	NNE	3°
CS-WX03b-1	AS030012b	653460	5146830	50 (2)	140°	46°	16°	NNE	3°
CS-WX08		653210	5136730	50 (13)	150°	31°	27°	E	6°
CS-WX11.5b		653270	5130050	30 (6)	117°	32°	19°	SE	1°
CS-X14		653720	5124940	30 (6)	013°	31°	22°	NE	2°
HV-C08		658440	5164930	50 (9)	118°	33°	13°	SE	12°
HV-C12b		657820	5157420	30 (3)	041°	16°	17°	NE	10°
HV-CD03.5-1		659240	5174020	30 (1)	266°	18°	17°	WNW	7°
HV-CD03.5-2		659240	5174020	30 (1)	268°	22°	19°	WNW	7°
HV-DE02		660920	5176790	60 (3)	359°	31°	10°	crest	0°
HV-DE14.5		660740	5152050	50 (8)	097°	30°	15°	E	7°
HV-F12b		664310	5156950	30 (1)	119°	22°	20°	ESE	4°
HV-FG07.5		665180	5166300	50 (10)	100°	32°	8°	SW	8°
HV-G04		665970	5173060	40 (6)	217°	27°	16°	S	3°
HV-G13b		666650	5155730	45	091°	32°	1°	NW	8°
HV-L11b		671920	5158950	40 (6)	034°	30°	20°	WNW	4°
HV-L13.5		672260	5154180	50 (2)	233°	25°	21°	WNW	7°
HV-M04.5		673710	5172070	40 (0)	338°	19°	20°	N	4°
HV-NP07		676600	5166460	25 (2)	075°	23°	19°	ENE	8°
HV-P11.5-1		678010	5158530	30 (1)	148°	15°	12°	NE	4°
HV-P11.5-2		678010	5158530	30 (4)	334°	17°	5°	NE	4°
HV-P11.5b	AS030015	677980	5157560	60 (12)	166°	29°	16°	S	10°
HV-Q01.5		680550	5177600	50 (11)	019°	36°	21°	NNE	2°
HV-Q14		680010	5153330	40 (6)	011°	27°	29°	NNE	1°
HV-R10-1		681890	5160980	30 (1)	188°	21°	12°	SSW	7°
HV-T12b		686420	5156400	40 (1)	240°	22°	27°	WNW	3°
HV-TV10		686920	5161120	45 (1)	274°	36°	11°	WNW	7°
HV-V05.5		687430	5169500	25 (1)	110°	19°	13°	ESE	10°
HV-V10		688320	5161070	40 (4)	130°	22°	32°	ESE	4°
HV-W03-2		689980	5174660	45 (4)	340°	29°	16°	N	4°
ND-A5b		653900	5142930	50 (2)	199°	39°	6°	E	4°
ND-BC03.5	AS030010	656780	5145790	50 (2)	353°	36°	16°	N	1°
ND-CD10	AS030054	658910	5133470	50 (21)	219°	33°	27°	SSW	8°
ND-D07.5	AS030055	659640	5137970	30 (5)	139°	26°	11°	NE	3°
ND-EF03.5	AS030060	662800	5146080	50 (9)	043°	41°	18°	SSE	1°
ND-FG06	AS030005	665370	5141210	60 (1)	042°	29°	18°	N	5°
ND-FG08.5-1	AS030048-1	664930	5136180	50 (6)	296°	25°	31°	WNW	4°
ND-FG08.5-2	AS030048-2	664930	5136180	30	282°	39°	5°	WNW	4°
ND-FG08.5-3	AS030048-3	664930	5136180	30	299°	21°	33°	WNW	4°

* Number of pebbles (a:b axis ratio ≥1.5) measured; figure in brackets is number of clasts interpreted as transverse, and deleted for calculation purposes.
† With respect to the horizontal.
‡ Regional slope, determined from the NTS topographic map.

fabric is primary, formed by the Younger Dryas glacier, while the granitic pebble content reflects the reworking of an older till (Seaman 2003). Results of work at site HV-P11.5 suggest that the realignment of till-fabric trends into a downslope-parallel orientation, as suggested by Lamothe (1992), may not be a factor for sites on moderate slopes in the order of 10°. However, realignment of the till-fabric plunge with respect to the underlying depositional surface may be a factor to consider in determining the ice-flow direction.

Hybrid tills, formed by one or more ice-flow events and then reworked by later events, have been documented in southwestern New Brunswick (Allard 2003; Allard and Pronk 2003). The pebble data for site HV-P11.5 clearly indicate reworking of underlying deformation till into the mixing zone of the overlying basal till by a northwest-flowing glacier. Our interpretation is that an older Wisconsinan till, derived initially from a source area to the west or northwest, was completely reworked by a northwestward-flowing glacier during the Younger Dryas.

Joe Lake

The only other reported radiocarbon dates of late glacial age from central New Brunswick are for sediment core from Joe Lake (Levesque et al. 1993a, 1993b; Mayle et al. 1993b; Mayle and Cwynar 1995a) in the southeastern part of the Tuadook Lake map area (Fig. 1). Two dates are reported: the lower date (11 110 ± 90 BP; TO-2303) is on the approximate lower boundary of a brief pre-Younger Dryas cooling interval (the Killarney Oscillation); and the upper date (10 800 ± 70; TO-2302) is on the approximate lower boundary of the Younger Dryas (Levesque et al. 1993a; Mayle et al. 1993b; Mayle and Cwynar 1995a). By extrapolating from these two dates, we can infer that the Joe Lake basin was deglaciated by approximately 11 500 BP. The Younger Dryas is marked by a decrease in sediment organic content, as determined by loss-on-ignition measurements; and by a lighter colour, which results from an increased minerogenic input arising from decreased vegetative productivity or increased inwash or solifluction (Levesque et al. 1993a).

From a glacial geology perspective, the significance of the Joe Lake site is that, while it clearly records effects of the Younger Dryas climatic cooling, its section contains no till, which implies that the lake basin was not overridden by Younger Dryas glaciers. This observation is compatible with the placement of the Younger Dryas ice margin (Fig. 1) suggested by Lamothe (1992, interpretation 2). In addition, the lack of glacial deposits in this lake basin offers circumstantial evidence against the eastward-flowing Younger Dryas glacier that Lamothe (1992, interpretation 1) postulated as the source of upper till in GSC trench 2009 at Todd Mountain (Fig. 2).

Joe Lake lies in an east–west-oriented valley that marks the surface expression of the Catamaran Fault (see Irrinki 1980, 1981; Crouse 1981) and is just north of the Todd

Mountain site. An ice cap that occupied the Miramichi Highlands and flowed eastward over several topographic obstacles to deposit 2+ m of till at an elevation of 380 m on Todd Mountain almost certainly would have completely occupied the Catamaran Fault valley situated 50 m lower in elevation. Therefore, non-glaciation of the Joe Lake basin adds to the weight of evidence favouring a Central Lowland centre — that being the Gaspereau Ice Centre — rather than a Miramichi Highlands centre for the Younger Dryas ice cap (Fig. 1).

Glacial Striae Trends

If the Younger Dryas ice cap occupied the western part of the New Brunswick Lowlands, as suggested by Lamothe (1992, interpretation 2), then expected ice-flow trends in the Todd Mountain area would be to the northwest. Farther south, a more westerly trend would be expected. Instead, most striae trends observed in central New Brunswick range from northeasterly clockwise through to southwesterly (Fig. 6; also see Seaman 1989a, 1989b). These striae trends probably relate to glaciation during the earlier Caledonia, Chignecto, and Bantalor phases of the Wisconsinan in New Brunswick (Rampton et al. 1984).

On the other hand, Chalmers (1895) reported northwest-trending striae from his sites 201 and 202 between Boiestown and Doaktown (Fig. 1) that he attributed to late-stage, local glaciers. Chalmers (1895, p. 72) described the striae at site 202 as having an indefinite trend of "N. 42° W., or the reverse," but reported those at site 201 as being "heavy" and accompanied by fine northerly trending striae. Striae of a similar northerly trend are plotted on Lamothe's (1990) Plate 1 for site 86-365 between Todd Mountain and Boiestown.

Although westerly trending striae are very rare in central New Brunswick (Fig. 7), Hamilton (1960) reported such striae from a site east of Grand Lake. Seaman (1989a) initially questioned the validity of this westerly trend, but fine striae accompanied by two wedge striae indicating westward flow have since been observed at a site between Fredericton and Grand Lake (A. Seaman, 1989 field notes, site 89-6). These fine striae cross-cut the striae, grooves, and rat-tails of the southerly trending Oromocto flow pattern of Rampton et al. (1984).

The above-noted striae all suggest a Gaspereau Ice Centre source during the Younger Dryas. However, northerly and northwesterly trending striae are common in eastern New Brunswick (Seaman 1989a, 1989b), including on the Tormentine Peninsula, where the Younger Dryas ice flow must have been southward. Also, west-trending striae have been observed in southwestern New Brunswick within the eskerine drainage basin of the Pennfield–Utopia and Pocologan delta complexes (see Seaman 1991, his Fig. 5). Therefore, while suggestive, none of the above striae can be definitively attributed to the Gaspereau Ice Centre or to the Younger Dryas.

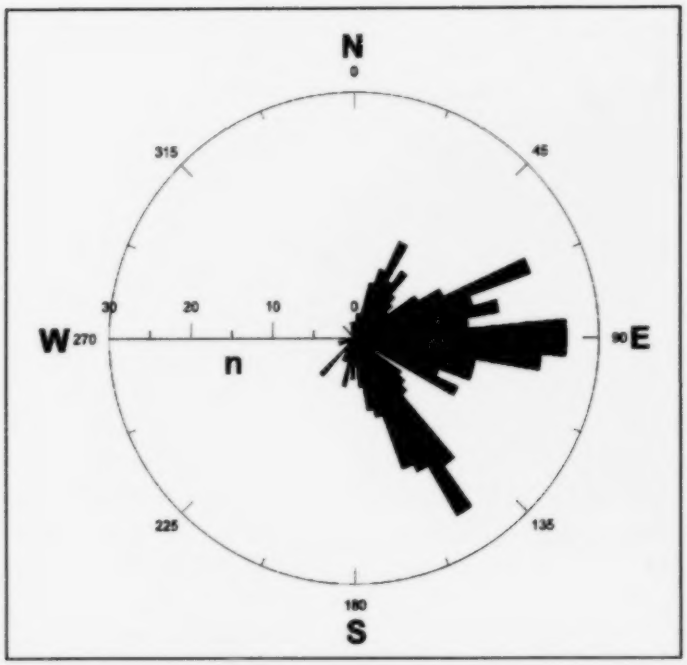

Figure 6. Summary plot of glacial striae trends in central New Brunswick covering the western quarter of the Moncton (NTS 21 I) map area and the eastern half of the Woodstock (NTS 21 J) map area.

Till-Fabric Trends

During the 2003 and 2004 field seasons, 50 till fabrics were measured at 45 sites in central New Brunswick (Table 1), specifically in the eastern part of the Coldstream (NTS 21 J/06) map area, the western part of the Napadogan (NTS 21 J/07) map area, and the Hayesville (NTS 21 J/10) map area (Fig. 1). The standard deviation around the mean for each till-fabric trend is on the order of 30° (see Table 1). At this level of precision, it can safely be said that the 50 till fabrics cover every compass direction (Fig. 7).

Sacré (1989) noted that most till-fabric trends in the Todd Mountain area were oriented downslope. About 60% of till-fabric trends listed in Table 1 lie within 1 standard deviation of the local topographic trend. Sacré (1989) suggested that downslope-trending till fabrics reflected either a reorientation of the fabric after glacial melting, or a change in ice-flow direction after dispersal train formation. Lamothe (1992) preferred the first interpretation; however, the following evidence from site HV-NP07 (Seaman et al. 2005) supports the second interpretation. The till section at this site lies only a few metres from a striated bedrock surface, which displays three sets of striae. The trend of the latest set lies within 2° of the mean trend of the till fabric. This suggests that the fabric was created by late-stage ice flow down the topographic slope. As noted by Seaman et al. (2005), the till fabric at a given site may be simply a snapshot of the last ice-flow trend to deform the till at that site. In many cases, this may be a late, downslope flow that involves limited transport distances. In contrast with the upper fabric at site HV-P11.5, the fabric at site HV-NP07 is strongly imbricated with an upslope plunge.

For till-fabric trends that are not oriented downslope, the trend can be used to infer regional ice-flow direction. However, the trends still cover all compass directions (Fig. 8). As well, north- to northwest-trending till fabrics are reported from southwestern New Brunswick (see Seaman 1991, his Fig. 8), making it difficult to definitively associate any fabric trends with the Younger Dryas.

An exception to the foregoing occurs in the Napadogan Brook area (Fig. 1), which lies in the Coldstream map area, just west of Lamothe's (1992) hypothetical Lowland ice cap, where till

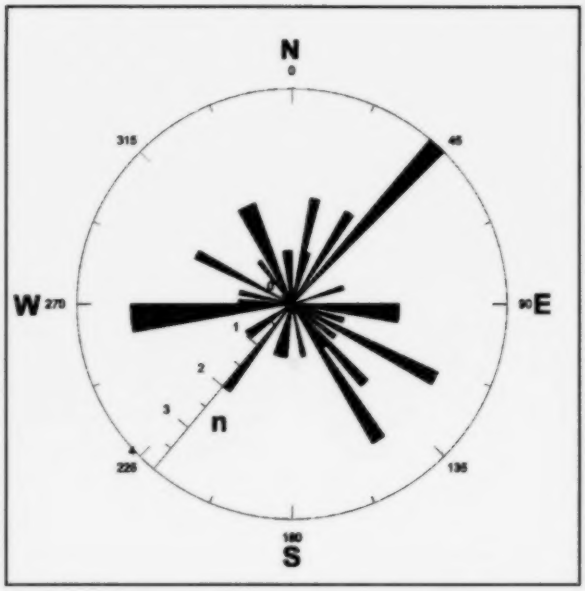

Figure 7. Summary plot of 50 till-fabric measurements from sites in the Coldstream (NTS 21 J/06), Napadogan (NTS 21 J/07), and Hayesville (NTS 21 J/10) map areas.

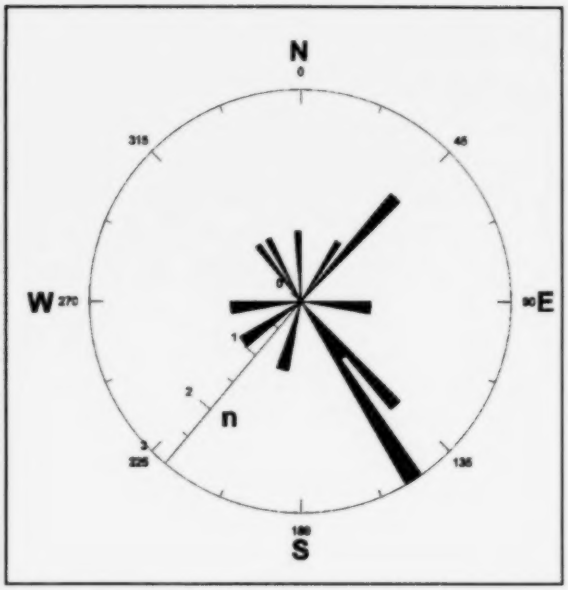

Figure 8. Subset of 20 till-fabric measurements from the data used for Figure 7 that do not lie within 1 standard deviation of the regional slope trend.

variability suggests ice from two different sources (Seaman 2003). The Napadogan Brook area is dominated by silty, sandy gravel till, with anomalous metal values derived from the Sisson Brook, Nashwaak, and Manzer Brook mineral occurrences (Seaman 2002a, 2003). Evidence suggests that till from sites CS-W08.5 and CS-W10b was probably last reworked by eastward- to northeastward-flowing glaciers (Fig. 9a, 9b). Till fabrics at these two sites may indicate late-stage Late Wisconsinan downslope flows (Table 1), as site CS-W10b lies within, and is part of, a southeast-trending dispersal train.

However, till at a third locale — site CS-W09 — is much finer than the area norm, being a gravelly, sandy mud that has metal values at background levels. The site (situated 1 km south of site CS-W08.5 and 2 km north of site CS-W10b) lies on a broad, flat crest and should be free of downslope influences. Till-fabric measurements at site CS-W09 (Fig. 9c; Table 1) indicate the till was deposited by a westward-flowing glacier. This westerly trend, coupled with the high silt-plus-clay content of the matrix, is consistent with a Younger Dryas ice cap flowing from the lowlands and suggests that the western margin of the Gaspereau Ice Centre may have been as far west as the Nashwaak Ridge.

New evidence from the nearby Hayesville map area, presented in the next section, gives additional support to the inferred presence of an active Gaspereau Ice Centre in central New Brunswick during the Younger Dryas.

Dispersal Trends from the Hayesville Map Area

Dominant glacial striae trends in the Hayesville map area (Fig. 1) are east to northeasterly (Seaman et al. 2005). Sacré (1989) and Lamothe (1992) reported that glacial dispersal in

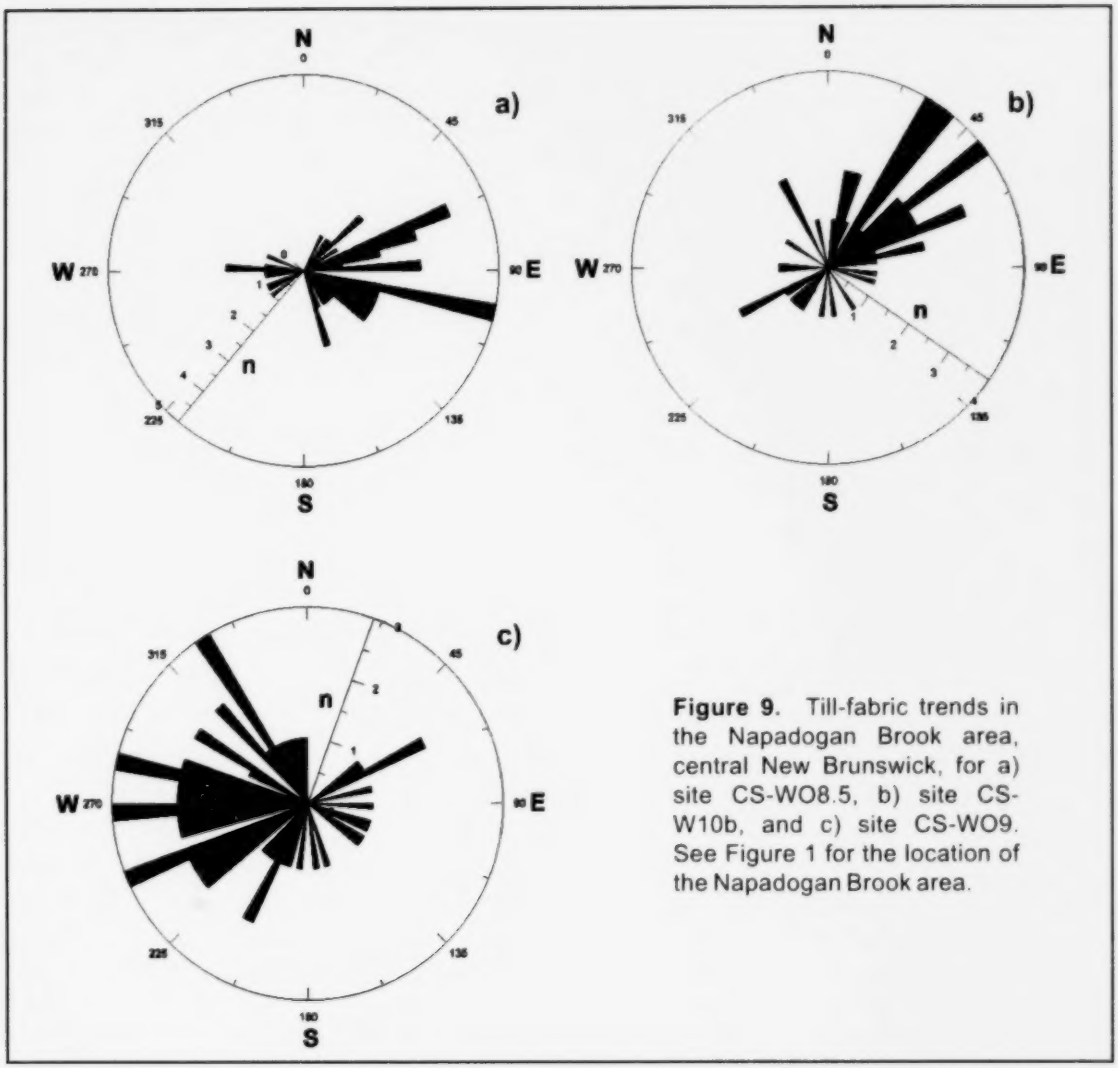

Figure 9. Till-fabric trends in the Napadogan Brook area, central New Brunswick, for a) site CS-WO8.5, b) site CS-W10b, and c) site CS-WO9. See Figure 1 for the location of the Napadogan Brook area.

this map area was also east-northeastward, with a magnitude of 1 km to 2 km. A recent till sampling project in the Hayesville map area using a 2 km grid (1 sample/4 km²) identified similar east-northeastward dispersal, primarily in the central and northeastern parts of the quadrant (Seaman et al. 2005). However, Seaman et al. (2005) also reported evidence of south-southeastward dispersal in the northern part of the map area, and north-northwestward dispersal in the southern part.

Lamothe (1990) reported a cross-cutting-striae site in the Hayesville map area, where early southeast ice flow was followed by east-northeast ice flow. A similar chronology was reported for southwestern New Brunswick (Seaman et al. 1993). On the basis of these reports, the south-southeast dispersal trends can be related to early ice-flow events, and the east-northeast dispersal trends to later flow events. Unlike the south-southeast-trending dispersal trains to the west, which taper to a point down-ice (see Seaman 2000), some of those of similar trend in the Hayesville map area appear truncated. The truncation occurs north of the Younger Dryas till site at GSC trench 2009 (Sacré 1989; Lamothe 1992), which lies

within the area of apparent northward-northwestward dispersal. This suggests that the north-northwestward ice flow is the latest in the area, and that truncation of the south-trending dispersal trains is due to the incursion of a Younger Dryas glacier from a New Brunswick Lowlands centre that reworked or buried previously deposited till and its associated dispersal trains.

Evidence for these glacial dispersal trends in the Hayesville map area comes from textural, pebble lithology, and geochemical data that were analyzed to determine distribution patterns. With the bedrock geology map of the area as a base layer (Table 2, Fig. 10), these textural, lithologic, and element distribution patterns (described below) were compiled using Surfer® 8 software to produce 10 distribution plots (see Fig. 11–20).

Table 2. Bedrock units of the Hayesville (NTS 21 J/10) map area, as illustrated in Figure 10. Unit descriptions were synthesized from Irrinki (1980, 1981), Crouse (1981), and Fyffe (1984).

System	Group	Formation	Map Unit	Lithology
Jurassic				Green diabase
Devonian		Buttermilk Brook Granite	LDBMfi	Pink biotite and muscovite-biotite granite; minor pink aplite, quartz-feldspar porphyry, and greisen
		Burnthill Granite	LDBUfi	Pink to red biotite granite, varying to quartz-feldspar porphyry; minor grey granite porphyry, pegmatite, and pink to reddish brown , quartz-feldspar porphyry dykes
		Dungarvon Granite	LDDGfi	Pink to red biotite and muscovite-biotite granite; minor pink to red aplite, quartz-feldspar porphyry, and greisen
		Sisters Brook Granite	LDSBfi	Pink biotite and muscovite-biotite granite; minor pink aplite, quartz-feldspar porphyry, and greisen
		Lost Lake Granite	EDLLfi	Pink, biotite-muscovite granite; grey to pink, massive to weakly foliated, biotite granite; grey to greenish grey foliated granodiorite; minor garnet-biotite-muscovite pegmatite
		Nashwaak Granite	EDNWfi	Grey to pink, biotite and hornblende-biotite granite and granodiorite
		Clearwater Brook Gabbro	EDCBmi	Green, massive to weakly foliated gabbro
Silurian	Kingsclear	Burtts Corner	LSBUc	Grey wacke, micaceous siltstone, and slate; minor pebble conglomerate
		Cross Creek	ESCRc	Grey wacke and slate; minor conglomerate and siltstone
Ordovician	Tetagouche			Grey wacke, siltstone and slate; minor grit and pebble conglomerate
		Hayden Lake	MOHLc	Green to maroon chert, and siltstone, interstratified with dark grey slate; quartzite pebble conglomerate interbedded with siltstone; minor felsic volcanic rocks
			MOHLmv	Grey to green mafic volcanic rocks; minor tuffaceous wacke and gabbro
			MOHLfv	Buff to greenish grey felsic volcanic rocks
		McKiel Lake Granite	MOMKfi	Pink, orange and red, foliated metagranite; pink- and grey-banded augen granite; pink to grey to green mylonitic equivalents; minor pegmatite
Cambro–Ordovician	Miramichi		COMc	Grey quartzite interbedded with siltstone, slate and phyllite; porphyroblastic psammite and pelite as contact metamorphic equivalents
	Trousers Lake Metamorphic Suite		COTLp	Quartzofeldspathic paragneiss interlayered with biotite-sillimanite paragneiss

within the area of apparent northward-northwestward dispersal. This suggests that the north-northwestward ice flow is the latest in the area, and that truncation of the south-trending dispersal trains is due to the incursion of a Younger Dryas glacier from a New Brunswick Lowlands centre that reworked or buried previously deposited till and its associated dispersal trains.

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System	Group	Formation	Map Unit	Lithology
Jurassic		Central Belt	JDcb	Green diabase
Devonian		Buttermilk Brook Granite	lDBMfi	Pink biotite and muscovite-biotite granite; minor pink aplite, quartz-feldspar porphyry, and greisen
		Burnthill Granite	lDBUfi	Pink to red biotite granite, varying to quartz-feldspar porphyry; minor grey granite porphyry, pegmatite, and pink to reddish brown , quartz-feldspar porphyry dykes
		Dungarvon Granite	lDDGfi	Pink to red biotite and muscovite-biotite granite; minor pink to red aplite, quartz-feldspar porphyry, and greisen
		Sisters Brook Granite	lDSBfi	Pink biotite and muscovite-biotite granite; minor pink aplite, quartz-feldspar porphyry, and greisen
		Lost Lake Granite	eDLLfi	Pink, biotite-muscovite granite; grey to pink, massive to weakly foliated, biotite granite; grey to greenish grey foliated granodiorite; minor garnet-biotite-muscovite pegmatite
		Nashwaak Granite	eDNWfi	Grey to pink, biotite and hornblende-biotite granite and granodiorite
		Clearwater Brook Gabbro	eDCBmi	Green, massive to weakly foliated gabbro
Silurian	Kingsclear	Burtts Corner	lSBUc	Grey wacke, micaceous siltstone, and slate; minor pebble conglomerate
		Cross Creek	eSCRc	Grey wacke and slate; minor conglomerate and siltstone
Ordovician	Tetagouche	Push and Be Damned	lOPBDc	Grey wacke, siltstone and slate; minor grit and pebble conglomerate
		Hayden Lake	MOHLc	Green to maroon chert, and siltstone, interstratified with dark grey slate; quartzite pebble conglomerate interbedded with siltstone; minor felsic volcanic rocks
			MOHLmv	Grey to green mafic volcanic rocks; minor tuffaceous wacke and gabbro
			MOHLfv	Buff to greenish grey felsic volcanic rocks
		McKiel Lake Granite	MOMKfi	Pink, orange and red, foliated metagranite; pink- and grey-banded augen granite; pink to grey to green mylonitic equivalents; minor pegmatite
Cambro–Ordovician	Miramichi		€OMc	Grey quartzite interbedded with siltstone, slate and phyllite; porphyroblastic psammite and pelite as contact metamorphic equivalents
	Trousers Lake Metamorphic Suite		€OTLp	Quartzofeldspathic paragneiss interlayered with biotite-sillimanite paragneiss

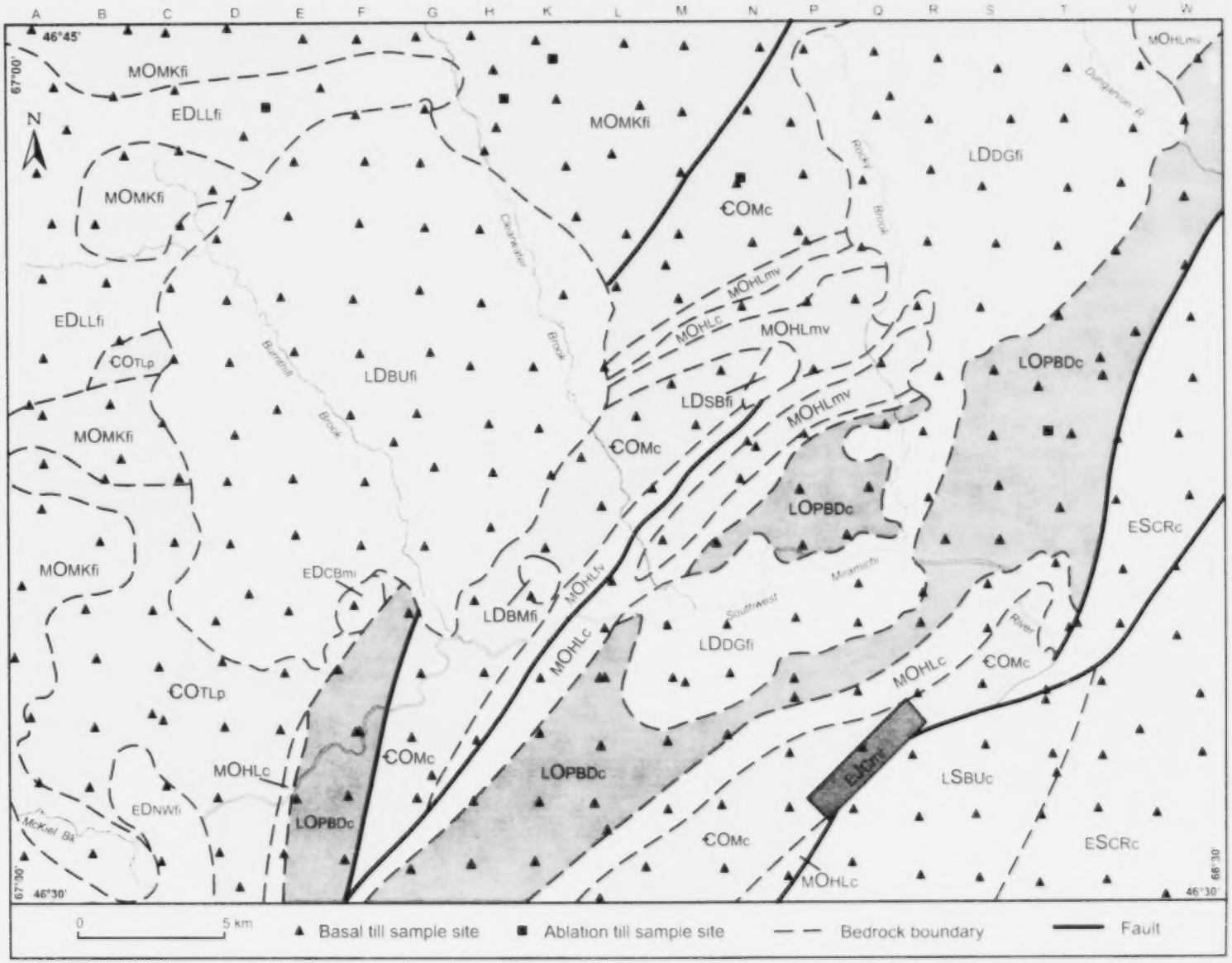
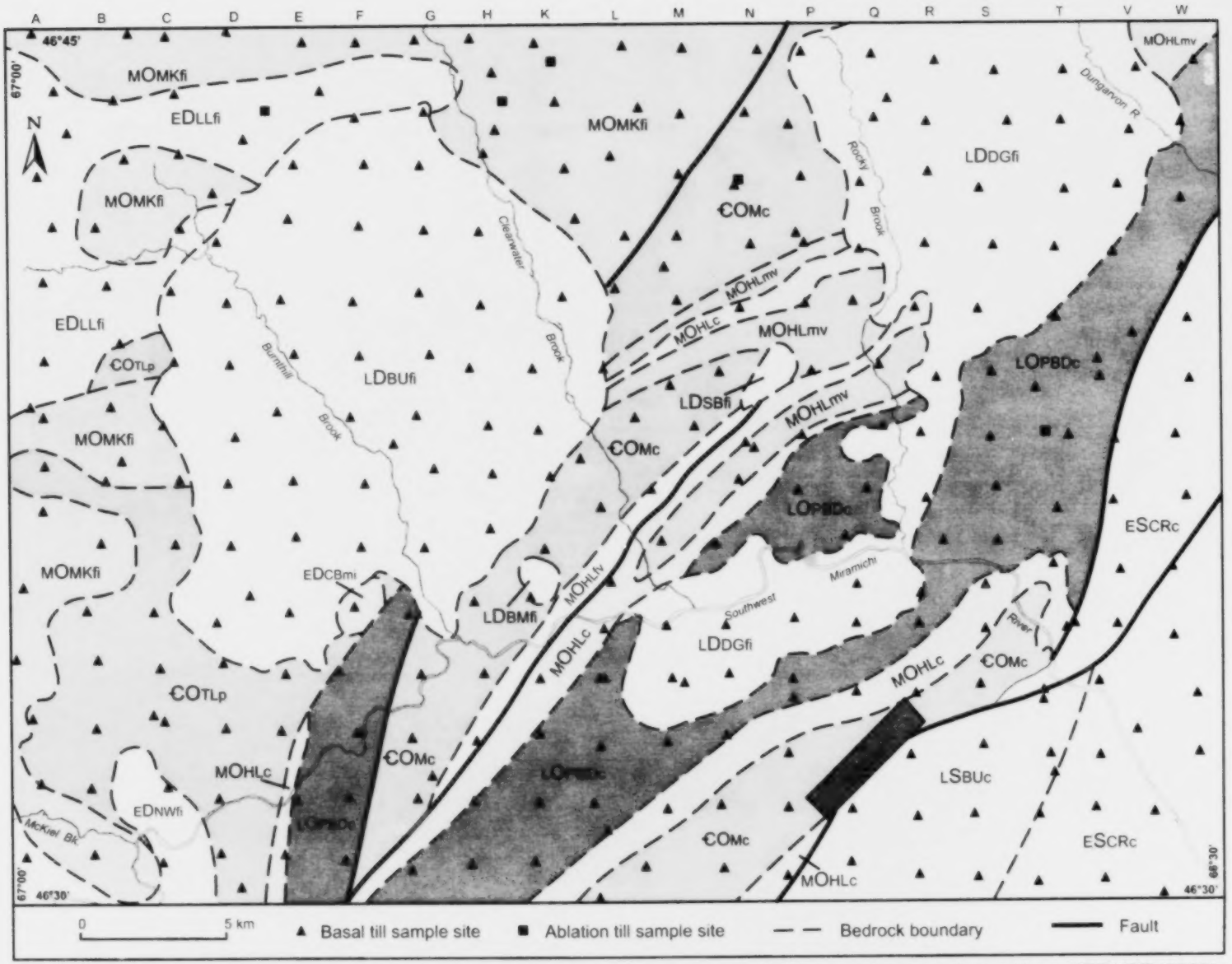


Figure 10. Bedrock geology and sample site locations in the Hayesville map area. Unit boundaries were derived from NBDNRE (2000), and unit labels have been updated to reflect current designations. See Table 2 for descriptions of units shown on this map.



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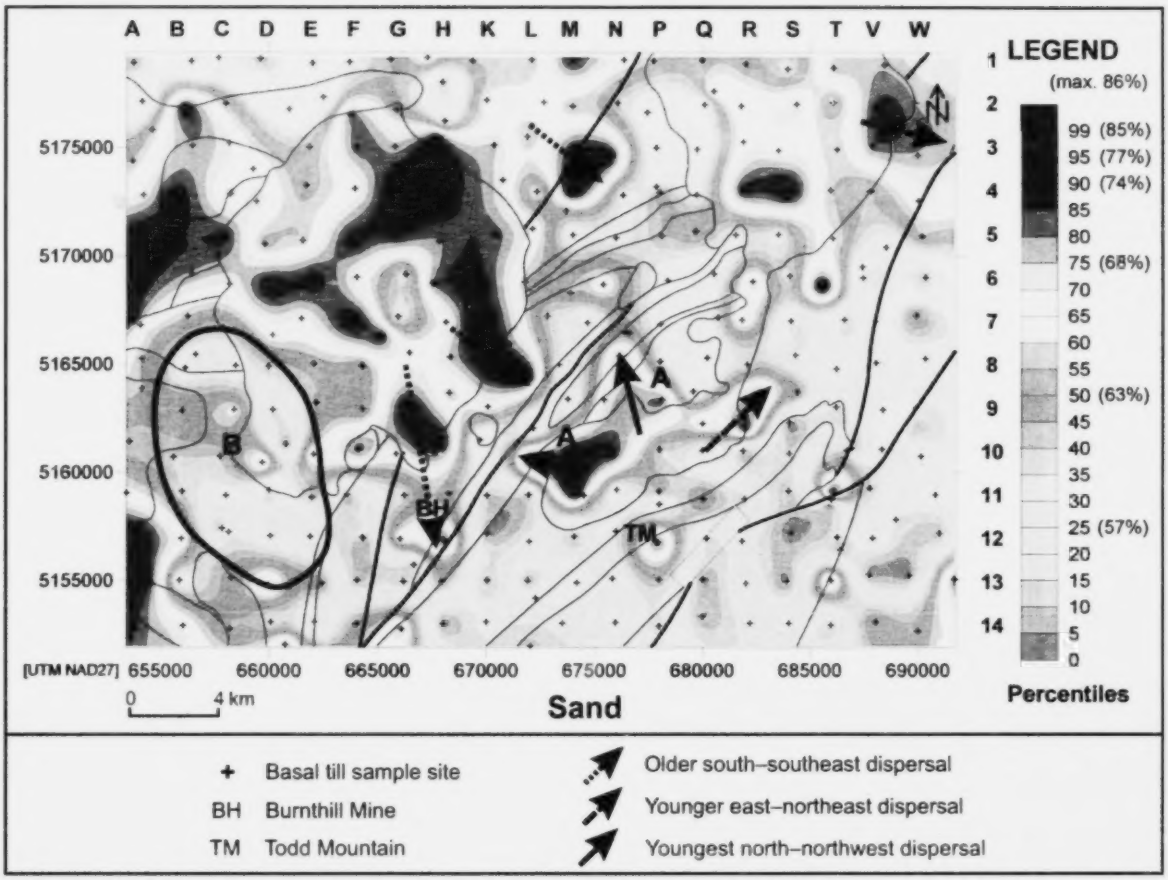

Figure 11. Sand distribution in basal till, Hayesville map area. Distribution plot shows possible northward dispersal from the Trout Lake lobe of the Dungarvon Granite (A) and comparatively low sand concentrations over parts of the Trousers Lake Metamorphic Suite and the Burnthill Granite (B). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Textural Distribution Patterns

Sand Plot

The plot for sand in the Hayesville map area shows an obvious association between granitic bedrock sources and basal till with a sand content greater than the 40th percentile (61%) (Fig. 11). South-southeastward dispersal of 2 km to 6 km is evident from the eastern part of the Burnthill Granite and from the McKiel Granite in the central part of the map area. The irregular areas of elevated sand content south of the Trout Lake lobe of the Dungarvon Granite (the Trout Lake Granite of MacLellan et al. 1990) may reflect a remnant of a south-southeast-trending dispersal train reworked by later ice-flow events. Eastward dispersal of similar magnitude is evident from the main body of the Dungarvon Granite, and from the eastern part of the Trout Lake lobe of that granite. In contrast, there appears to be north-northwestward dispersal on the order of 1 km to 2 km from the central and western parts of the Trout Lake lobe. Also, immediately west of the former Burnthill Mine, a belt of low sand till extends across the Trousers Lake Metamorphic Suite and onto the southwestern part of the Burnthill Granite.

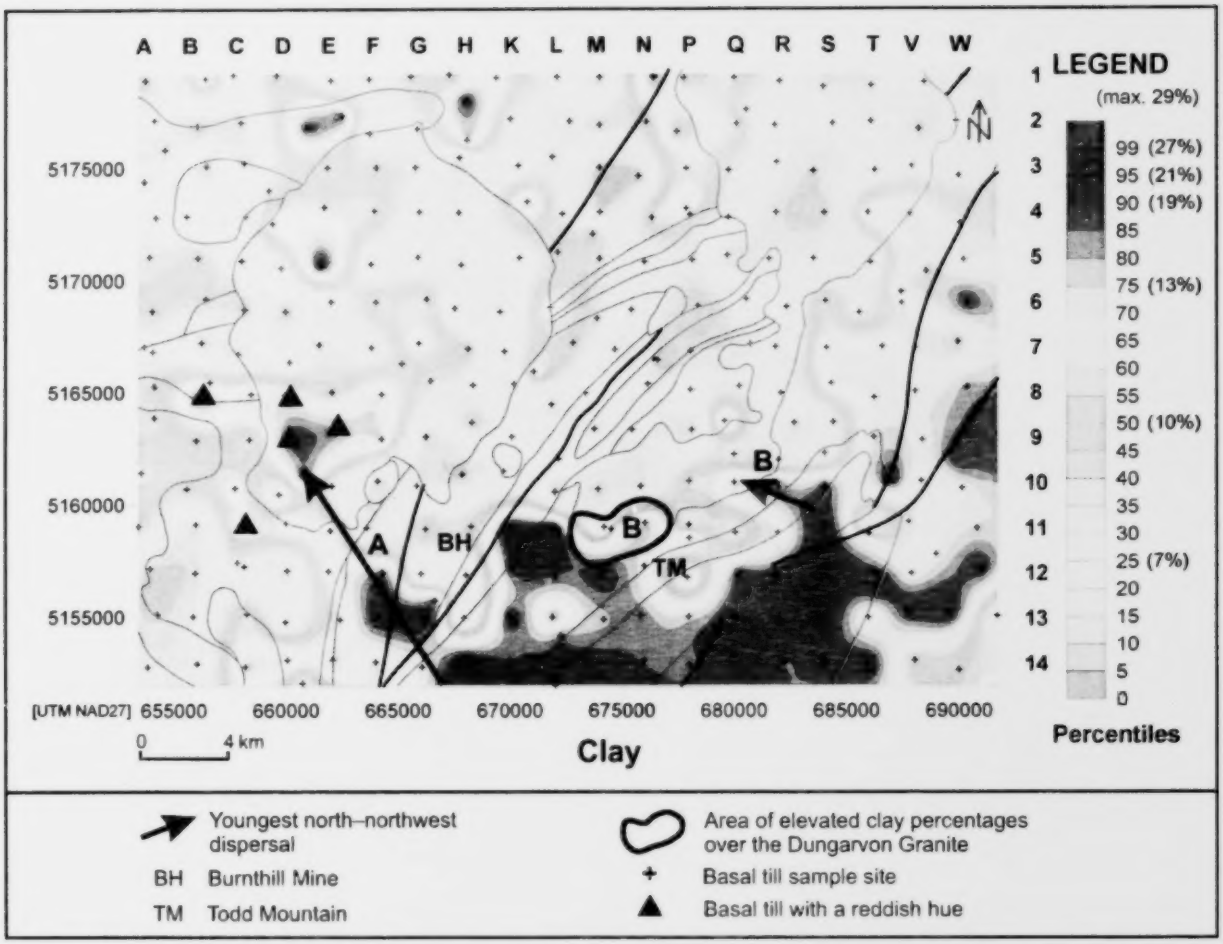

Figure 12. Clay distribution in basal till, Hayesville map area. The distribution plot shows northwestward dispersal onto the Trouser Lake Metamorphic Complex paragneiss and the southern part of the Burnthill Granite (A) and the Trout Lake lobe of the Dungarvon Granite (B). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Clay Plot

The plot for clay is almost the reverse of that for sand (Fig. 12). Of note are clay values up to the 75th percentile level (13%) over the southern part of the Trout Lake lobe, and the belt of elevated clay values that extends northwestward onto the southwestern part of the Burnthill Granite. Till at five sites in the latter area (HV-B08b, HV-C11, HV-D08, HV-D09, and HV-E09) is reddish brown to yellowish red (Munsell® soil colour hue of 5YR) (Seaman et al. 2005). Till of similar hue is associated with Carboniferous red clastic sedimentary rocks in the Coldstream map area (Fig. 1; Seaman 2002b), indicating that Hayesville tills of this colour may also be derived from Carboniferous bedrock. The extent of these north-northwest-trending dispersal trains suggests that the Younger Dryas glacier extended northward to approximately the position of the Southwest Miramichi River, except in the southwestern part of the Hayesville map area, where a lobe may have extended northwestward across the river, up the valleys of several of its southeast-flowing tributaries.

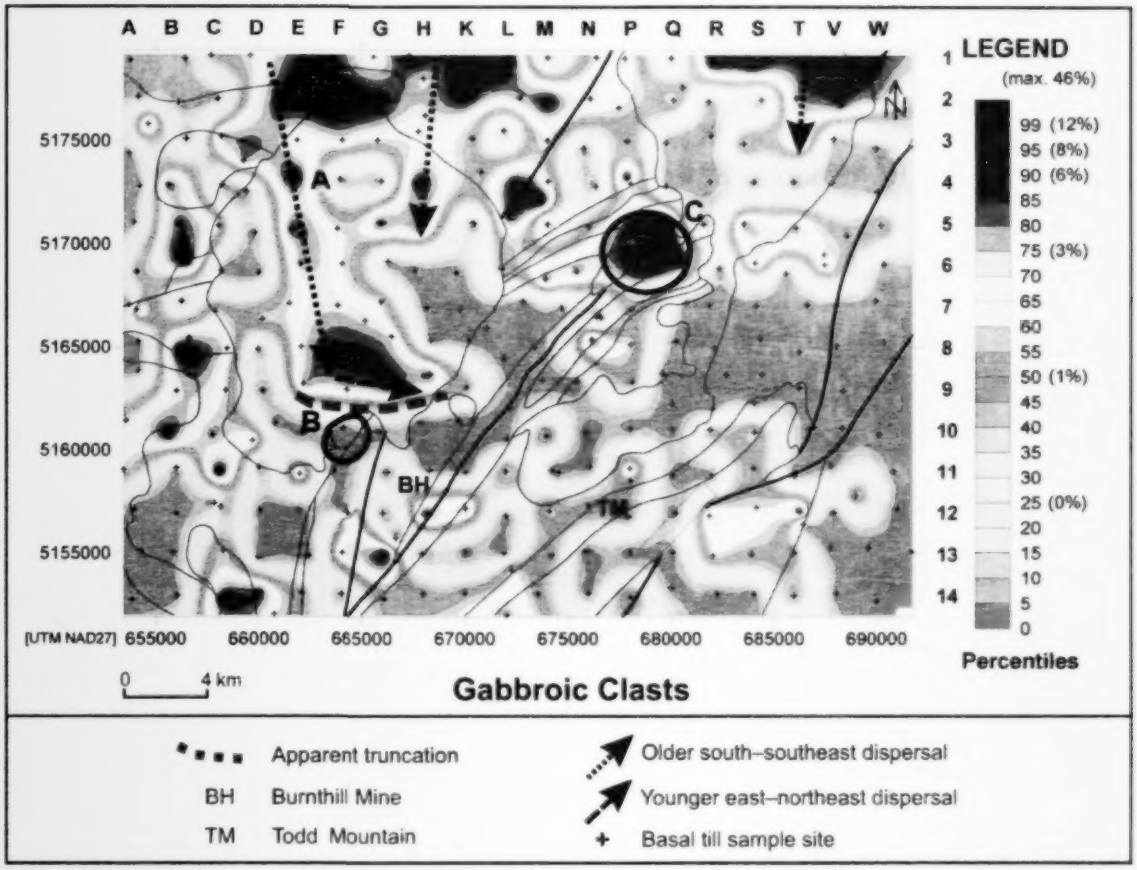

Figure 13. Distribution of gabbroic clasts in basal till, Hayesville map area, illustrating the southward-trending dispersal train from the Clearwater Brook Gabbro to the north. The dispersal train crosses most of the Burnthill Granite (A), and appears to be truncated north of a small area of gabbro (B) that shows no obvious appearance in the dispersal patterns. The extreme anomaly in the central map area (C) correlates with a mafic phase of the Hayden Lake Formation. Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Lithologic Distribution Patterns

Gabbroic Clasts

The distribution plot of gabbroic clasts in basal till for the Hayesville map area shows an extreme high over a mafic phase of the Hayden Lake Formation (Fig. 13). Elevated gabbroic clast contents also occur at sites to the northwest and northeast. The former sites probably represent southward and southeastward dispersal from the Clearwater Gabbro, north of the Catamaran Fault (Crouse 1981). The latter sites likely represent similar dispersal from a small body of Ordovician gabbro that also lies north of the fault (Irrinki 1981).

The south-trending gabbro dispersal train across the Burnthill Granite is about 16 km long. It almost completely crosses the granite but terminates abruptly just north of another small gabbroic area (Fig. 10) that has no obvious representation on the gabbro distribution diagram (Fig. 13). The train's termination shape suggests it is a palimpsest dispersal train (Parent et al. 1996) formed by the earlier, southerly trending ice-flow event and reworked in

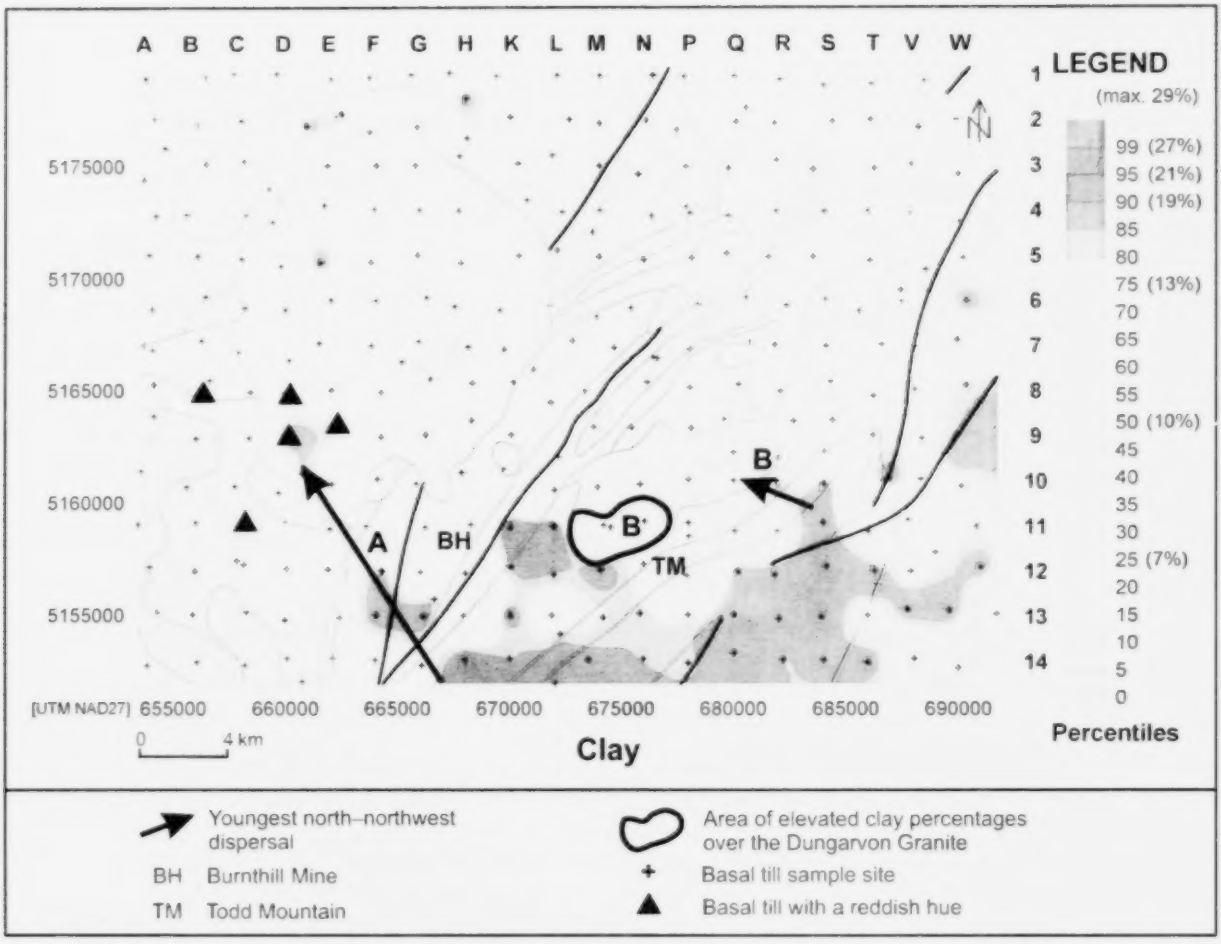

Figure 12. Clay distribution in basal till, Hayesville map area. The distribution plot shows northwestward dispersal onto the Trouser Lake Metamorphic Complex paragneiss and the southern part of the Burnthill Granite (A) and the Trout Lake lobe of the Dungarvon Granite (B). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Clay Plot

The plot for clay is almost the reverse of that for sand (Fig. 12). Of note are clay values up to the 75th percentile level (13%) over the southern part of the Trout Lake lobe, and the belt of elevated clay values that extends northwestward onto the southwestern part of the Burnthill Granite. Till at five sites in the latter area (HV-B08b, HV-C11, HV-D08, HV-D09, and HV-E09) is reddish brown to yellowish red (Munsell soil colour hue of 5YR) (Seaman et al. 2005). Till of similar hue is associated with Carboniferous red clastic sedimentary rocks in the Coldstream map area (Fig. 1; Seaman 2002b), indicating that Hayesville tills of this colour may also be derived from Carboniferous bedrock. The extent of these north-northwest-trending dispersal trains suggests that the Younger Dryas glacier extended northward to approximately the position of the Southwest Miramichi River, except in the southwestern part of the Hayesville map area, where a lobe may have extended northwestward across the river, up the valleys of several of its southeast-flowing tributaries.

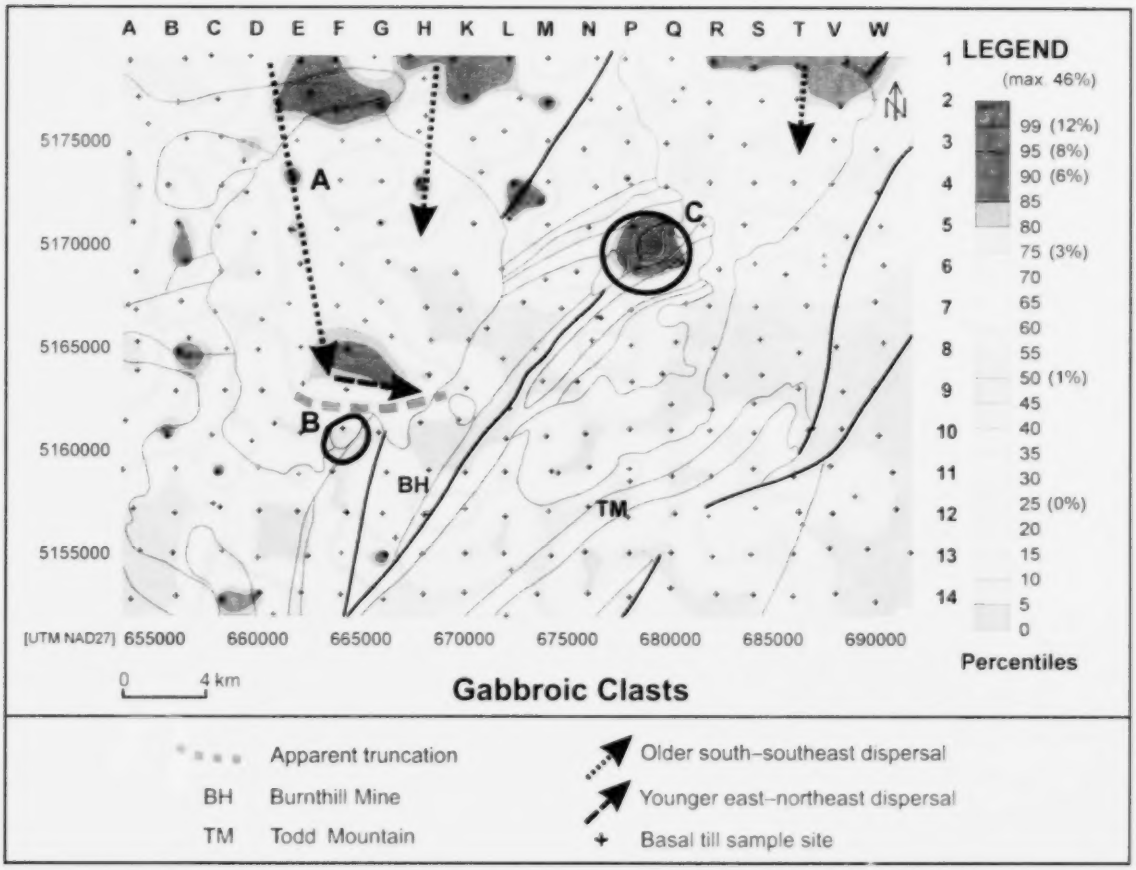

Figure 13. Distribution of gabbroic clasts in basal till, Hayesville map area, illustrating the southward-trending dispersal train from the Clearwater Brook Gabbro to the north. The dispersal train crosses most of the Burnthill Granite (A), and appears to be truncated north of a small area of gabbro (B) that shows no obvious appearance in the dispersal patterns. The extreme anomaly in the central map area (C) correlates with a mafic phase of the Hayden Lake Formation. Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Lithologic Distribution Patterns

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The south-trending gabbro dispersal train across the Burnthill Granite is about 16 km long. It almost completely crosses the granite but terminates abruptly just north of another small gabbroic area (Fig. 10) that has no obvious representation on the gabbro distribution diagram (Fig. 13). The train's termination shape suggests it is a palimpsest dispersal train (Parent et al. 1996) formed by the earlier, southerly trending ice-flow event and reworked in

part by the later, eastward ice flow. The abrupt termination also suggests this dispersal train was truncated by a later glacial event, possibly by the eastward ice flow. However, the termination location of the south-trending dispersal train lies adjacent to the northwest-trending low sand/elevated clay dispersal train noted above, which suggests that formation of the latter dispersal train may have been responsible for the truncation. As well, the small gabbro body has no obvious eastward dispersal south of the truncation of the south-trending dispersal train; this further supports the case for truncation by the northwest-trending dispersal train.

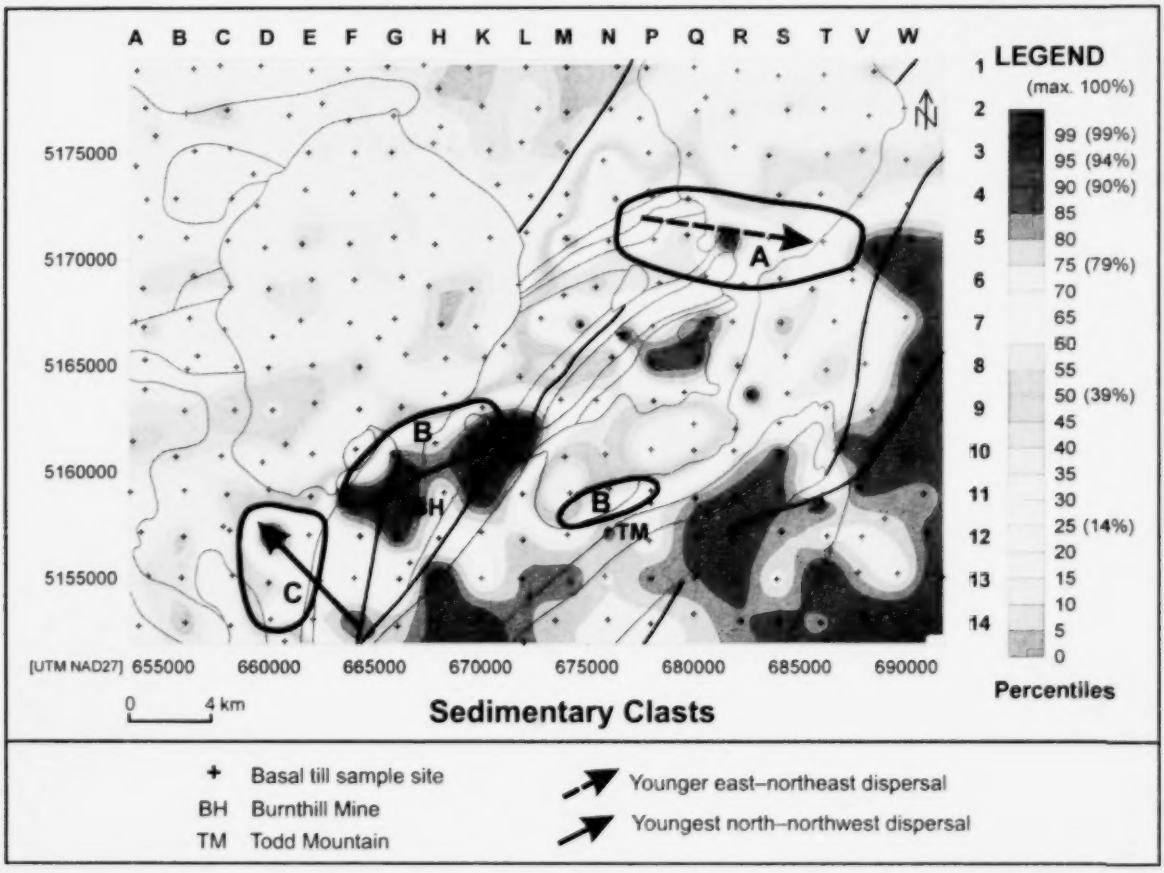


Figure 14. Distribution of clastic sedimentary clasts in basal till, Hayesville map area. Distribution plot shows 1) the eastward dispersal of clasts from the Miramichi and Tetagouche groups over the central part of the Dungarvon Granite (A), 2) the apparent northward dispersal onto the southernmost parts of the Burnthill and Dungarvon granites (B), and 3) the apparent northwestward dispersal onto the Trousers Lake Metamorphic Suite (C). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Sedimentary and Granitic Clasts

The distribution plot of sedimentary clasts in the Hayesville map area shows a strong association with areas underlain by sedimentary rocks (Fig. 14). The distribution plot of granitic-type clasts (which include gneissic clasts from the Trousers Lake Metamorphic Complex) is almost a mirror image of the sedimentary clasts plot (Fig. 15). Several characteristics of these plots are noteworthy.

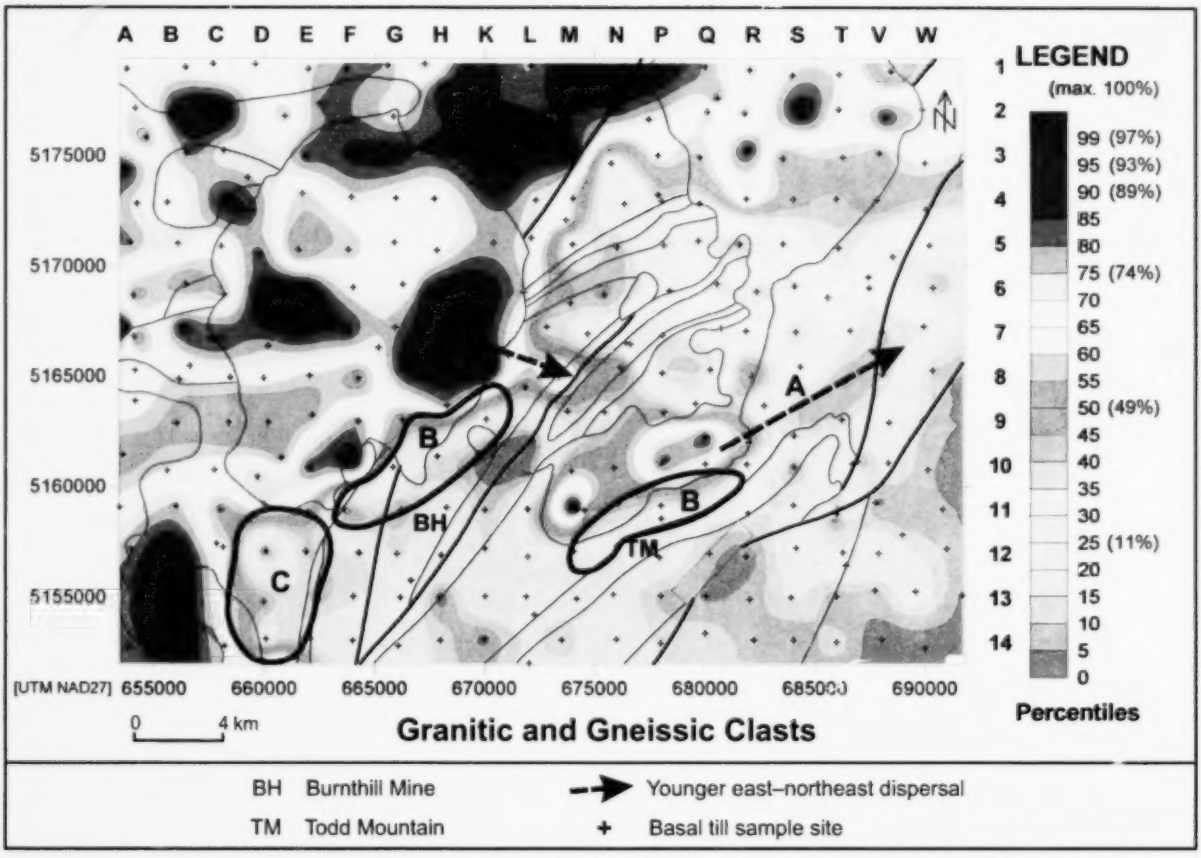


Figure 15. Distribution of granitic and gneissic clasts in basal till, Hayesville map area. Distribution plot shows 1) a diffuse east-northeast-trending dispersal train from the Trout Lake lobe of the Dungarvon Granite (A), 2) the absence of significant southward or southeastward dispersal from the latter area or the southern part of the Burnthill Granite (B), and 3) the absence over part of the Trousers Lake Metamorphic Suite of eastward dispersal from the Nashwaak Granite (C). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

- In the the northeastern part of the map area, sedimentary clasts are dispersed eastward across the lower part of the main body of the Dungarvon Granite and its Rocky Brook lobe (the Rocky Brook Granite of MacLellan et al. 1990), with attendant low granitic clast content.
- A dispersal train of granitic clasts, at about the 50th percentile level (49%), extends east-northeast for 8 km from the Trout Lake lobe at the southern end of the Dungarvon Granite.
- There is no apparent southward or southeastward dispersal of granitic clasts from the Burnthill Granite or the Trout Lake lobe, despite the obvious southward dispersal of gabbroic clasts across the Burnthill Granite.
- Sedimentary clasts appear to be dispersed northward about 1 km onto the Trout Lake lobe, and over the Buttermilk Brook Granite and the adjacent southeastern point of the Burnthill Granite.
- In the southwestern corner of the map area, there is no apparent eastward dispersal of clasts from the Nashwaak Granite or the Trousers Lake Metamorphic Complex. Instead, sedimentary clasts appear to be dispersed northwestward over the southern part of the Trousers Lake Metamorphic Complex.

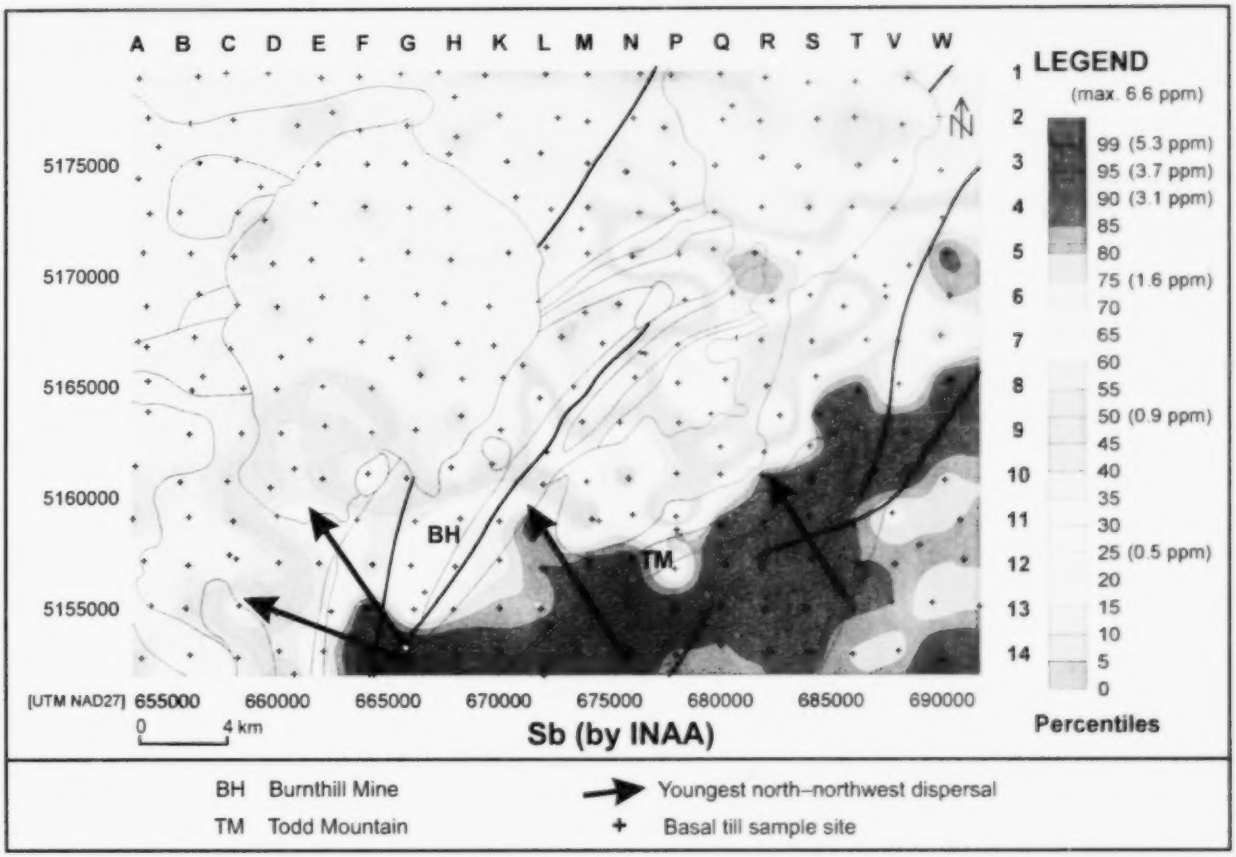


Figure 16. Sb distribution in basal till, Hayesville map area. Distribution plot shows possible northwestward dispersal from the Silurian Kingsclear Group in the southeast, or from its fault contact with the Miramichi Group. Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Element Distribution Patterns

Antimony

The Sb values are high in basal till over the southern limb of the syncline intruded by the Dungarvon Granite, but not over the northern limb (Fig. 16). The same clastic sedimentary rock units occur in both limbs, although the northern limb includes mappable areas of felsic to mafic volcanic rocks that are not mapped in the southern limb.

It is unlikely that the volcanic rocks, which cover significantly less area than do the sedimentary rocks, could account for the low Sb values in till over the northern limb. Most of the highest Sb values occur in till near the fault contact between the southern limb of the syncline and the adjacent Silurian sedimentary rocks of the Fredericton Trough, and over the latter rocks, to the southeast. If this area is the source of the Sb, the element distribution pattern suggests northwestward dispersal of several kilometres onto the southern limb of the syncline.

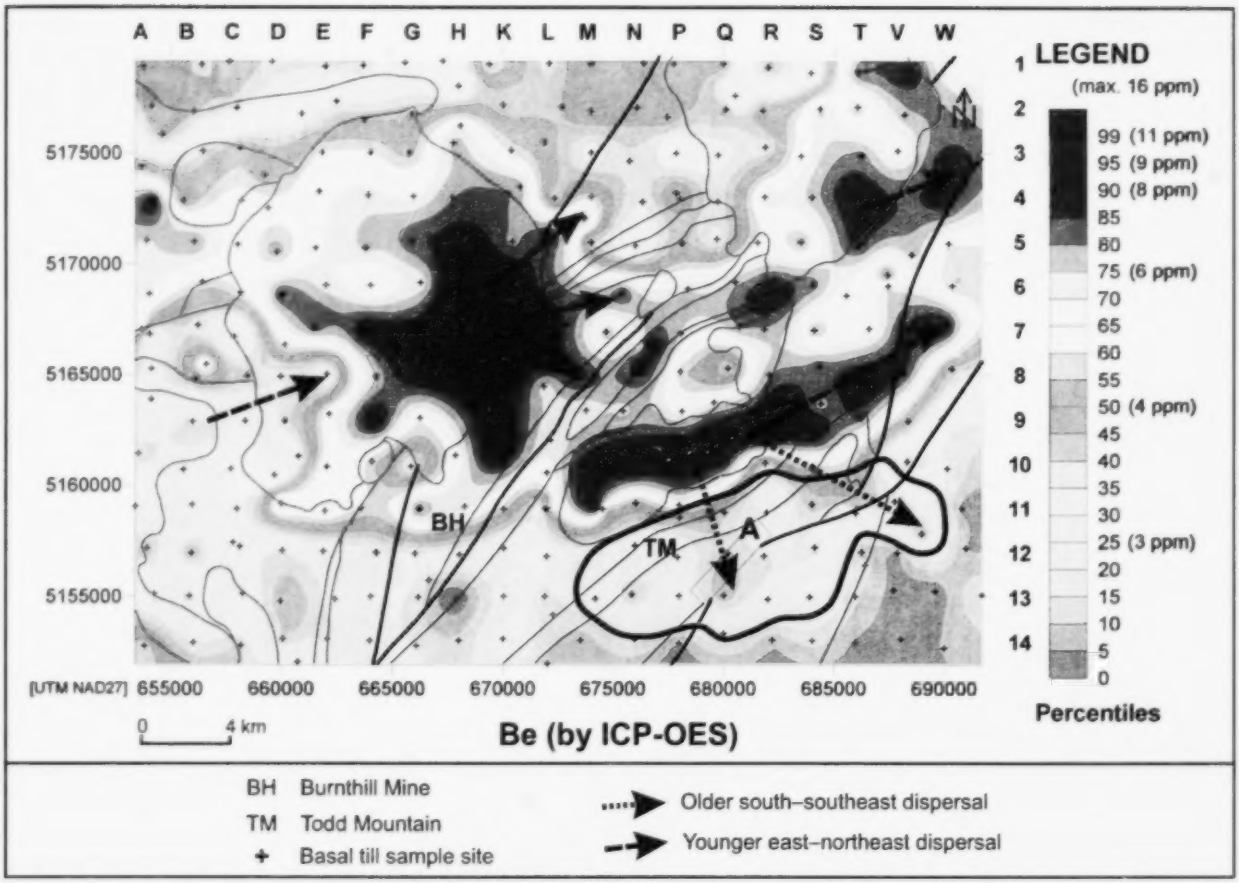


Figure 17. Be distribution in basal till, Hayesville map area. Distribution plot shows short northeastward to southeastward dispersal trains from the Burnthill Granite, and longer east-northeastward dispersal trains from the Dungarvon Granite. The area of till with values in the 20th to 40th percentile range (3 ppm) situated south and southeast of the Trout Lake lobe of the Dungarvon Granite (A) may reflect old dispersal trains reworked by later north-northwestward glacier flow. Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Beryllium

There is an obvious association between elevated Be values in basal till and the Late Devonian granites of the Hayesville map area (Fig. 17). Northeastward to southeastward dispersal on the order of 2 km to 4 km is evident in the area of the Burnthill Granite, both in elevated values east of the pluton and in low values over its western margins. Dispersal distances appear greater in the eastern part of the map area, where the western half of the Dungarvon Granite is covered by low-Be till, and where a 10 km dispersal train appears to extend east-northeast from the Trout Lake lobe. No significant southward dispersal of similar magnitude has been observed in the latter area. However, the broad area south of the Trout Lake lobe with Be in the 25th to 50th percentile range (34 ppm) may reflect a remnant of a south-trending dispersal train reworked by northwestward ice flow during the Younger Dryas. Generally similar patterns (not illustrated) are evident for Rb, Sn, Ta, Th, and U. Eu also exhibits a similar pattern, but in the negative sense: that is, a negative Eu anomaly is associated with these Late Devonian granites.

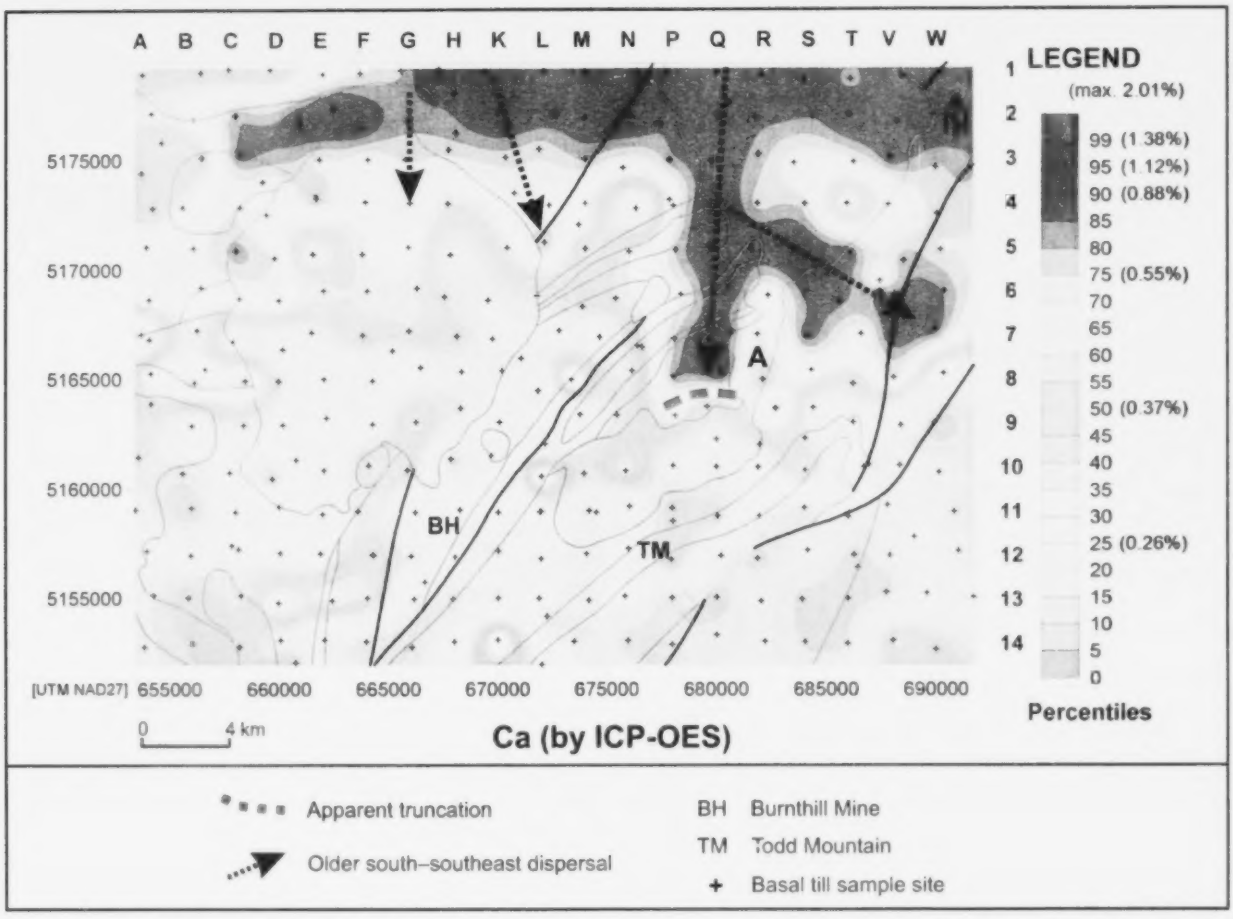


Figure 18. Ca distribution in basal till, Hayesville map area. Distribution plot shows southward and southeastward dispersal, with apparent truncation in the east-central part of the area (A). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Calcium

High Ca values occur in basal till across the northern part of the Hayesville map area (Fig. 18). The distribution pattern suggests that the values are related to the granodioritic phase of the Lost Lake Granite in the northwest (see Crouse 1981) and to mafic plutonic rocks north of the map boundary (see Irrinki 1980, 1981; Crouse 1981).

There are several southerly trending lobes of elevated Ca values that indicate southward dispersal. Most significant is the one in the northeast quarter of the map area. This lobe crosses several bedrock units over a distance of 10 km, then appears to be abruptly truncated near the Southwest Miramichi River in the centre of the map area. A second dispersal train trending southeasterly appears to cross the middle of the southerly trending one. Its irregular, lobate shape suggests that it is an older dispersal train that subsequently was reworked in part by a later, southward ice flow.

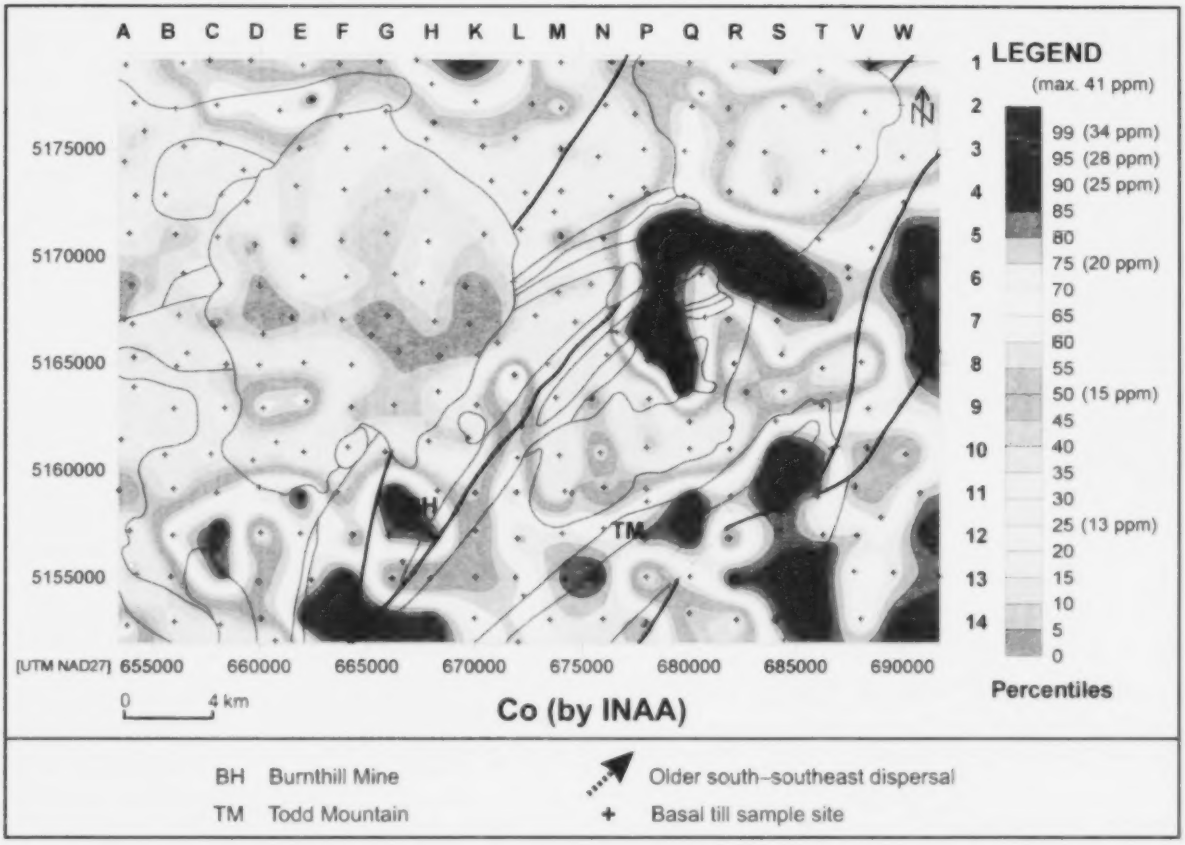


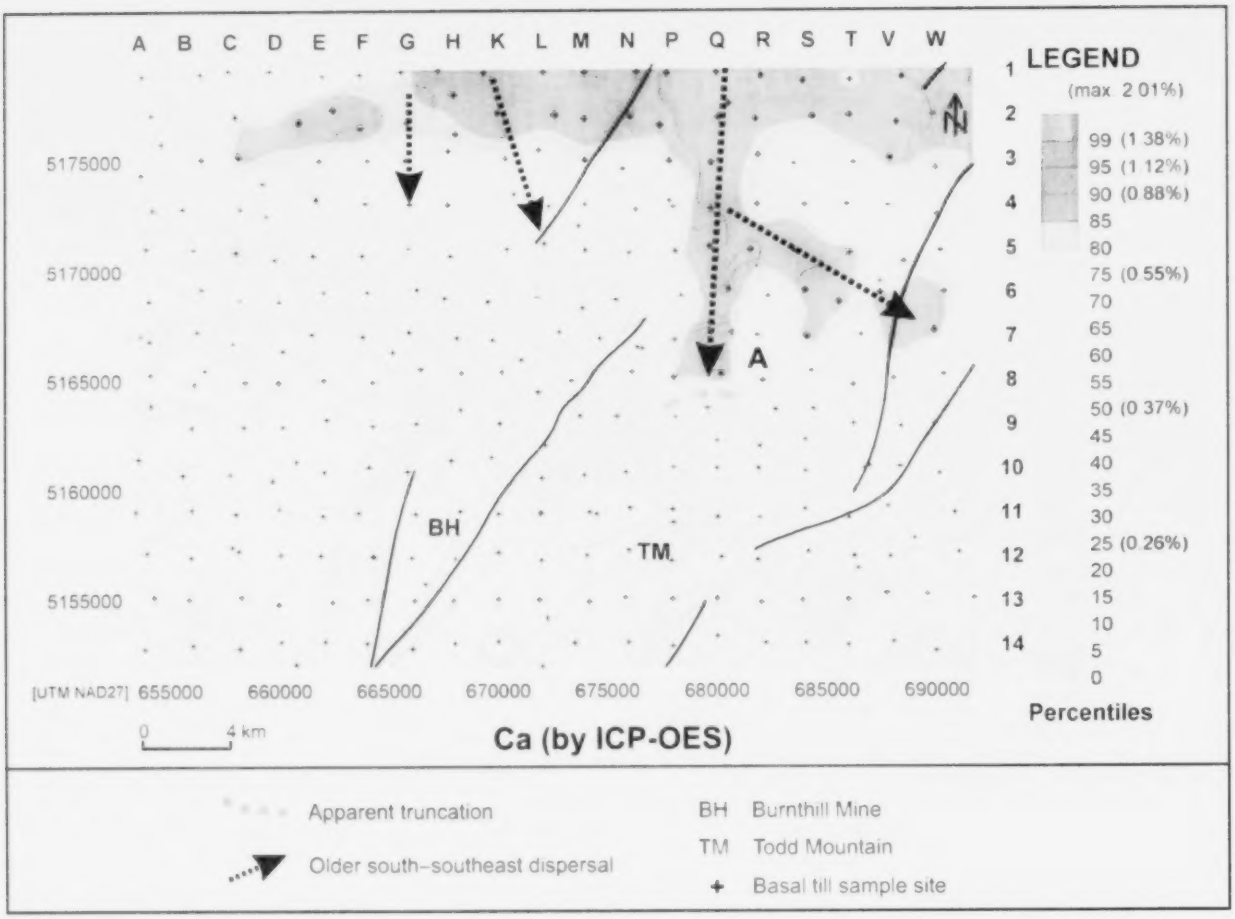
Figure 19. Co distribution in basal till, Hayesville map area. Distribution plot shows southward and southeastward dispersal from a mafic phase of the Hayden Lake Formation, with a dispersal pattern similar to that for Ca in this area (see Fig. 18). Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

Cobalt

A Co anomaly in basal till is associated with the area of mafic volcanic rocks of the Hayden Lake Formation between the main body of the Dungarvon Granite and the Sisters Brook Granite to the southwest (Fig. 19). Two distinct dispersal trains extend from this area. One trends southeast, parallel to but slightly south of the southeast-trending Ca dispersal train noted above. The other train trends southerly, coincident with the similar Ca dispersal train, and appears to terminate in the same area. Generally similar patterns (not illustrated) are present for Cr, Cu, Mg, Mn, and Ni.

Sodium

All rocks in the Hayesville map area have either a marine sedimentary, or an igneous, origin. Marine sedimentary rocks of the Miramichi and Tetagouche groups in this map area are comparable in age and depositional environment to rocks of the Woodstock and Meductic groups in west-central New Brunswick. Basal till derived from these rocks to the west (Fig. 20) typically has Na values greater than the 40th percentile (1.29%). In contrast, till derived



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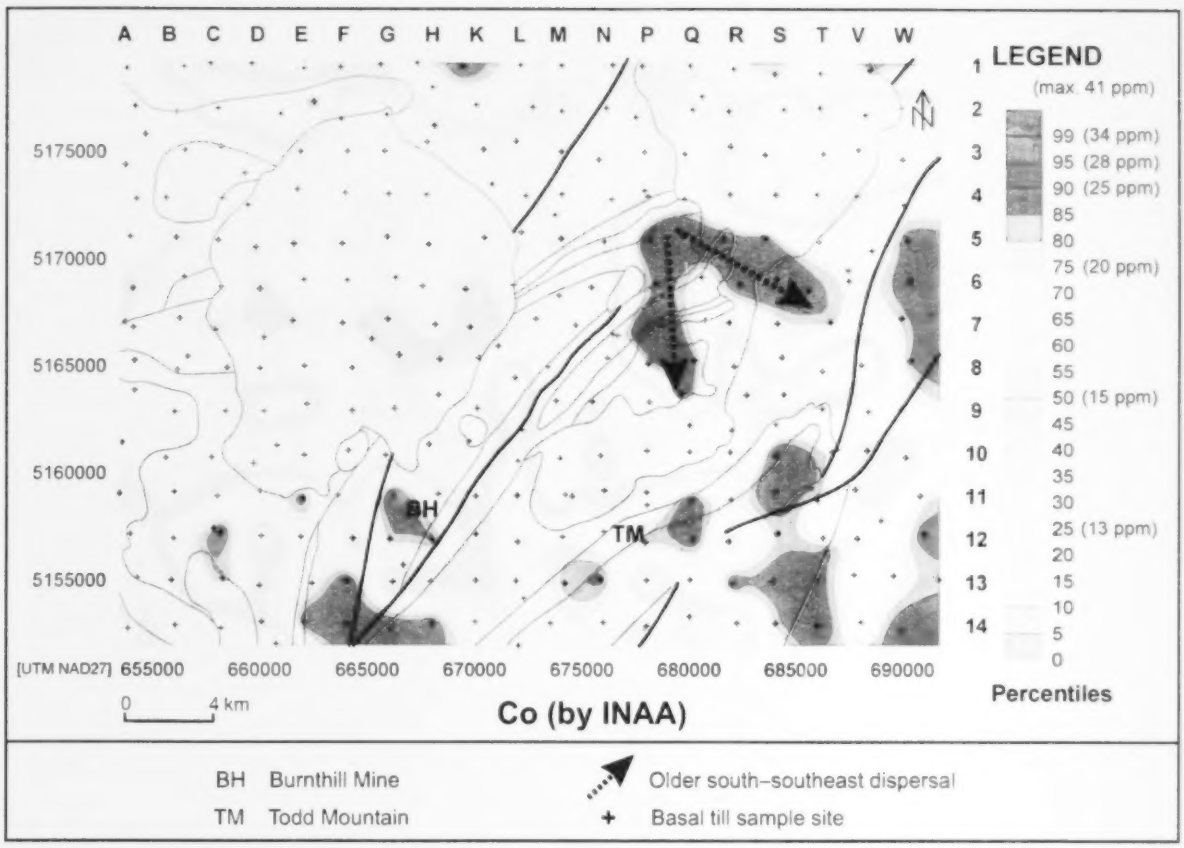

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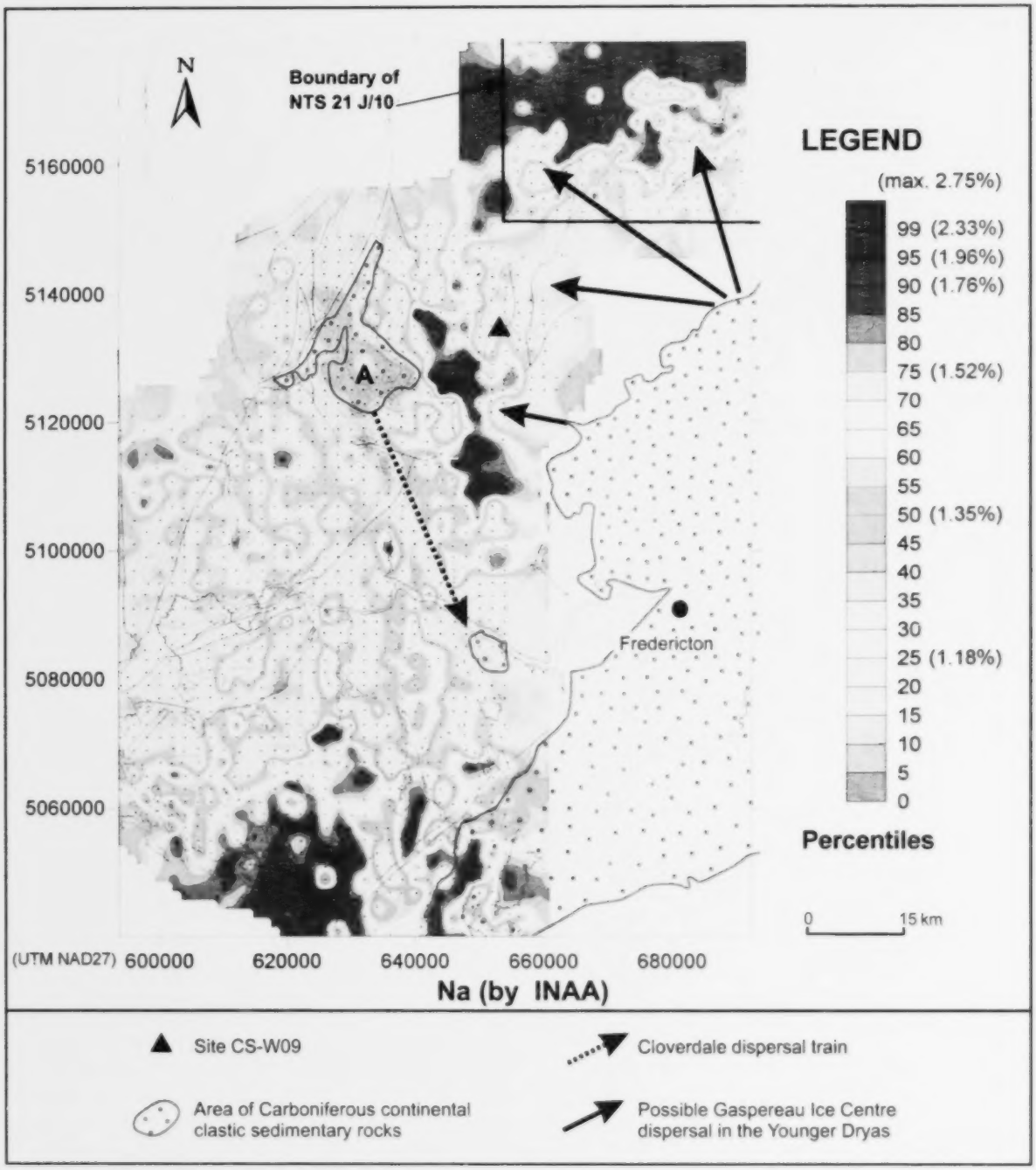

Figure 20. Na distribution in west-central New Brunswick and part of central New Brunswick. Distribution plot shows 1) the long, south-southeast-trending dispersal train of low-Na till extending from the area of Carboniferous continental clastic sedimentary rocks in the Cloverdale district (A), and 2) the possible westward to northwestward dispersal from similar rocks of the New Brunswick Lowlands in central New Brunswick. Fine lines underlying the distribution plot represent the boundaries of bedrock geology units (see Fig. 10).

from continental clastic sedimentary rocks of the Carboniferous Carlisle and Mountain View formations is characterized by Na values below the 20th percentile (1.14%) and forms a distinct south-southeast-trending dispersal train (Fig. 20). For reference, the Na content of most felsic to intermediate igneous rocks is >2.00%, while that of most mafic igneous rocks is >1.25% (Heier and Billings 1970a). The Na content in sedimentary rocks is varied, depending on the mineralogy and depositional environment (Heier and Billings 1970b). The Na content in metamorphic rocks is little changed from the original source rock (Heier and Billings 1970c).

Till in the southern part of the Hayesville map area and in the Napadogan Brook area to the southwest (Fig. 1) has Na values that typically are less than the 40th percentile (Fig. 20). In some areas, they are even low enough to be comparable to till derived from the Carlisle and Mountain View formations. This suggests an influx of glacial debris from the nearest continental clastic sedimentary rocks, which would be those of the Carboniferous Pictou Group of the New Brunswick Lowlands to the southeast. The intermediate Na values in some of these areas may reflect reworking and incorporation of previously deposited till of original Miramichi Highlands source.

Discussion

The organic-bearing clayey diamicton in GSC trench 2009 at Todd Mountain — whether it is within a single till unit as suggested by Sacré (1989) or at the base of an upper till overlying a lower till as suggested by Lamothe (1992) — strongly suggests that an active glacier was present in central New Brunswick during the Younger Dryas. Lamothe's assessment (Lamothe 1992, interpretation 1) of this till as having been deposited by an eastward-flowing glacier is inconsistent with 1) the northwest-trending till fabrics reported from the site, 2) an apparent northerly oriented glaciotectonic structure at the north end of the trench, and 3) his own hypothetical interpretation (Lamothe 1992, interpretation 2) of a Younger Dryas ice centre over the New Brunswick Lowlands and the eastern part of the Miramichi Highlands. The absence of easterly trending dispersal patterns in most of the southern part of the Hayesville map area is further evidence against Younger Dryas glaciation by eastward-flowing ice from a highlands source.

New dispersal trends identified in the Hayesville map area suggest that the southern part of the area, including the Todd Mountain site (GSC trench 2009), was subject to different glacial conditions than was the northern part. East- to south-trending dispersal trains in the northern part are comparable to dispersal trains in west-central New Brunswick (Seaman 2000). However, unlike those latter dispersal trains, the ones in the Hayesville map area are comparatively short, and some seem truncated at their southern ends. This truncation appears to coincide with the northern limit of the north-northwest-trending dispersal patterns characteristic of the southern part of the map area, suggesting that the latter dispersal trend

is associated with a younger event. The presence of dated till at Todd Mountain indicates that this younger event can be attributed to a Younger Dryas-age glacier. The north-northwestward dispersal suggests that the Gaspereau Ice Centre was the active glacier in this area at that time. The absence of obvious east- to south-trending dispersal trains in the southern part of the map area is attributed to either reworking or burial of older tills by Younger Dryas glaciation.

The hypothetical 11 000 to 11 500 BP ice margin (Fig. 1) proposed by Lamothe (1992, interpretation 2, his Fig. 22), appears to provide a realistic approximation of the extent of the Gaspereau Ice Centre during the Younger Dryas, although topographic influences probably caused its margin to be lobate and irregular rather than smoothly curved. The Todd Mountain site lies just inside the northwestern margin of this ice centre. Conversely, Joe Lake to the north and Killarney Lake to the southwest (at Fredericton) lie beyond this hypothetical ice margin. Both lakes record climatic deterioration associated with the Younger Dryas but were not occupied by glacial ice. The till at site CS-W09 in the Napadogan Brook area (Fig. 1), with its high clay content, low metal values, and westerly oriented till fabric, suggests that this area may mark the westernmost limit of the Gaspereau Ice Centre, only a short distance west of Lamothe's (1992) proposed ice margin.

Lamothe (1992) indicated that the northern, eastern, and southern margins of the Younger Dryas glacier are uncertain. To the south, it probably extended into the Grand Lake basin. As noted above, the lower date from the Joe Lake core is older than 11 000 BP. The lowest reported core dates from other lakes in southwestern New Brunswick, including Killarney Lake, are also older than 11 000 BP (Levesque et al. 1993a, 1994). In contrast, a date of 8160 ± 76 BP (BGS-77) is reported by Rampton et al. (1984) for organic materials from near the base of a bog between Grand Lake and Fredericton. The younger date suggests late deglaciation of this area, which is consistent with the presence of the Gaspereau Ice Centre during the Younger Dryas.

In the southeast, the Gaspereau Ice Centre may have been contiguous with an ice cap over the Caledonian Highlands. However, contiguity with the Younger Dryas ice cap over Prince Edward Island is problematic. Rampton et al. (1984) reported dates of 12 600 ± 140 BP (GSC-3284) and 11 830 ± 950 BP (S-1969(B)) for a whale vertebra collected from marine clay beneath a sand veneer at Saint-Louis-de-Kent on the east coast of New Brunswick (Fig. 1). The stratigraphy as described suggests this area was not overridden by a Younger Dryas glacier. However, if the Gaspereau Ice Centre resulted from initial outgrowth from an ice cap centred over the Caledonian Highlands, there may have been continuous ice across the southern part of the New Brunswick Lowlands that did not extend as far north as Saint-Louis-de-Kent (Fig. 1). Further research will be required to evaluate the extent of Younger Dryas glaciers in the New Brunswick Lowlands, including a program of bog or lake coring and dating of basal organic sediments.

CONCLUSIONS

The northern part of the Hayesville (NTS 21 J/10) map area exhibits the same easterly through southerly dispersal directions that have been reported from west-central New Brunswick. In contrast, evidence strongly suggests that the southern part of the map area was glaciated by a glacier flowing from an ice centre to the southeast. The presence of dated Younger Dryas till at Todd Mountain with a probable northwest-trending till fabric indicates that the glaciation took place during the Younger Dryas Chronozone.

Younger Dryas glaciation in the southern part of the Hayesville map area is consistent with ice flow from the Gaspereau Ice Centre. The ice centre was originally proposed by Rampton et al. (1984) as a local ice cap that was active during earlier phases of the Late Wisconsinan glaciation of New Brunswick. However, recognition that the main Late Wisconsinan glaciation was related to a single glacier of regional extent, rather than to coalescing local glaciers (Seaman 2004), and that New Brunswick was not ice-free by 12 000 BP, changes the conceptual model within which the Gaspereau Ice Centre was originally defined.

The Gaspereau Ice Centre consequently is redefined here as the ice cap that occupied the western part of the Central Lowland of the New Brunswick Lowlands, and parts of adjacent physiographic areas, during the Younger Dryas. Further research will be required to ascertain the extent of the Gaspereau Ice Centre, and to determine whether the ice centre was fully independent, or contiguous with other ice centres over the Caledonian Highlands and Prince Edward Island.

ACKNOWLEDGEMENTS

Troy Piercy assisted with the till-fabric measurement project in the Coldstream, Napadogan, and Hayesville map areas in 2003. The till sampling project in the Hayesville map area in 2004 was carried out in conjunction with Serge Allard, Rex Boldon, and Toon Pronk. Assistance with the Hayesville project fieldwork was provided by Will Gilmore, Peter Leonard, and Angela Page.

Craig Evans took on the daunting task of performing the lithologic identification of 292 bags of pebbles collected during the Hayesville sampling project as his B.Sc. Honours thesis project at the Department of Geology, University of New Brunswick. Jill Craig compiled raw data from the Evans study into the lithological groups used in this paper.

The figures were prepared or finalized by Terry Leonard. The paper was reviewed by Toon Pronk and Les Fyffe.

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THE NEPISIGUIT A DEPOSIT: A VENT-PROXIMAL, VOLCANOGENIC MASSIVE-SULPHIDE DEPOSIT, BATHURST MINING CAMP, NORTHEASTERN NEW BRUNSWICK

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The Nepisiguit A volcanogenic massive-sulphide deposit is located in the central portion of the Bathurst Mining Camp, northeastern New Brunswick. It is hosted within the largely volcanic sequence of the Spruce Lake Formation in the lower part of the California Lake Group. Tholeiitic basalt in the immediate hanging wall of the deposit is overlain by a thick section of felsic crystal tuff and tuffaceous sedimentary rocks, typical of the Spruce Lake Formation. Lithogeochemical data indicate that the interlayered alkali basalt and sedimentary rocks assigned to the Spruce Lake Formation in the footwall of the deposit are similar to those in the coeval Canoe Landing Lake Formation.

The Nepisiguit A deposit consists of two subparallel, easterly thickening, massive-sulphide lenses that strike approximately east-northeast and dip about 80°N. Intersections of 8.93% Zn over 9.75 m have been reported from the larger, northern lens. The deposit as a whole contains an inferred resource of 1.54 million tonnes grading 2.8% Zn, 0.6% Pb, and 0.4% Cu. Massive sulphides in the deposit are dominated by pyrite but have subordinate pyrrhotite, sphalerite, galena, and chalcopyrite.

Stringer-sulphide veins cutting the footwall sedimentary rocks in the western part of the deposit suggest the presence of a feeder zone at depth in that area. Magmatic sulphur (as indicated by low $\delta^{34}S$ values) likely played an important role in forming the stringer sulphides (average $\delta^{34}S$ = 2.16‰) and proximal stratiform massive sulphides (average $\delta^{34}S$ = 4.94‰) that occur in the western part of the deposit. Conversely, a stronger input from seawater is suggested by the higher sulphur values (average $\delta^{34}S$ = 7.7‰) that characterize stratiform massive sulphides in the more distal, eastern part of the deposit.

Le gîte sulfure massif volcanogène de Nepisiguit A est situé dans la partie centrale du Camp minier de Bathurst, dans le Nord-Est du Nouveau-Brunswick. Il se trouve à

l'intérieur de la séquence fortement volcanique de la Formation de Spruce Lake, dans la partie inférieure du groupe du lac California. Le basalte tholéiitique dans la lèvre supérieure immédiate du gîte est recouvert d'une section épaisse de tuf cristal-felsique et de roches sédimentaires tufacées, caractéristiques de la Formation de Spruce Lake. Des données lithogéochimiques révèlent que les roches sédimentaires et le basalte alcalin interlités rattachés à la Formation de Spruce Lake dans la lèvre inférieure du gîte, sont semblables à ceux de la Formation du même âge de Canot Landing Lake.

Le gîte Nepisiguit A est constitué de deux lentilles de sulfures massifs subparallèles qui s'épaississent en allant vers l'est, sont orientées à peu près en direction est-nord-est et s'inclinent d'environ 80 degrés vers le nord. Des recoupements de 8,93% de Zn sur 9,75 mètres ont été signalés de la lentille de taille supérieure septentrionale. Le gîte dans son ensemble abrite des ressources inférées de 1,54 million de tonnes d'une teneur de 0,4% de Cu, de 0,6% de Pb et de 2,8% de Zn.

La pyrite prédomine dans les sulfures massifs à l'intérieur du gîte, mais ce dernier renferme aussi des minéraux secondaires tels que de la pyrrhotite, de la sphalérite, de la galène et de la chalcopyrite. Les filonnets de sulfures recouvrant les roches sédimentaires de la lèvre inférieure dans la partie occidentale du gîte laissent supposer la présence d'une zone nourricière en profondeur dans ce secteur. Le soufre magmatique (comme le révèlent les faibles concentrations de $\delta^{34}S$) ont probablement joué un rôle important dans la formation des filonnets de sulfures (en moyenne $\delta^{34}S$ = 2,16‰) et des sulfures massifs stratiformes proximaux (en moyenne $\delta^{34}S$ = 4,94‰) présents dans la partie occidentale du gîte. Inversement, les concentrations de soufre plus prononcées (en moyenne $\delta^{34}S$ = 7,7‰) caractérisant les sulfures massifs stratiformes de la partie orientale plus distale du gîte permettent de supposer un apport plus marqué d'eau de mer.

INTRODUCTION

The Bathurst Mining Camp (BMC) is a world-class, volcanogenic massive-sulphide district (Fig. 1). Since the discovery of the Brunswick No. 6 orebody in 1952 (Belland 1992), the BMC has produced more than 134 million tonnes of ore grading 7.82% Zn, 3.05% Pb, 0.46% Cu, and 91.5 g/t Ag (McCutcheon et al. 2003). Except for the Canoe Landing Lake deposit (Fig. 2), all 47 known volcanogenic massive-sulphide deposits in the BMC are spatially associated with felsic volcanic rocks (McCutcheon et al. 2001).

The Nepisiguit A deposit is located in the BMC, 16 km south-southwest of the Brunswick No. 12 deposit (Fig. 1, 2), close to the Nepisiguit B and Nepisiguit C deposits. The Nepisiguit A, B, and C deposits are catalogued as Unique Reference Number 241, 240, and 239, respectively, in New Brunswick's computerized mineral occurrence database (see Rose and Johnson 1990). All three deposits were discovered in 1956 by Kennco Exploration during the initial exploration rush in the BMC (Stevenson 1955; Stevenson and Brooks 1956). The

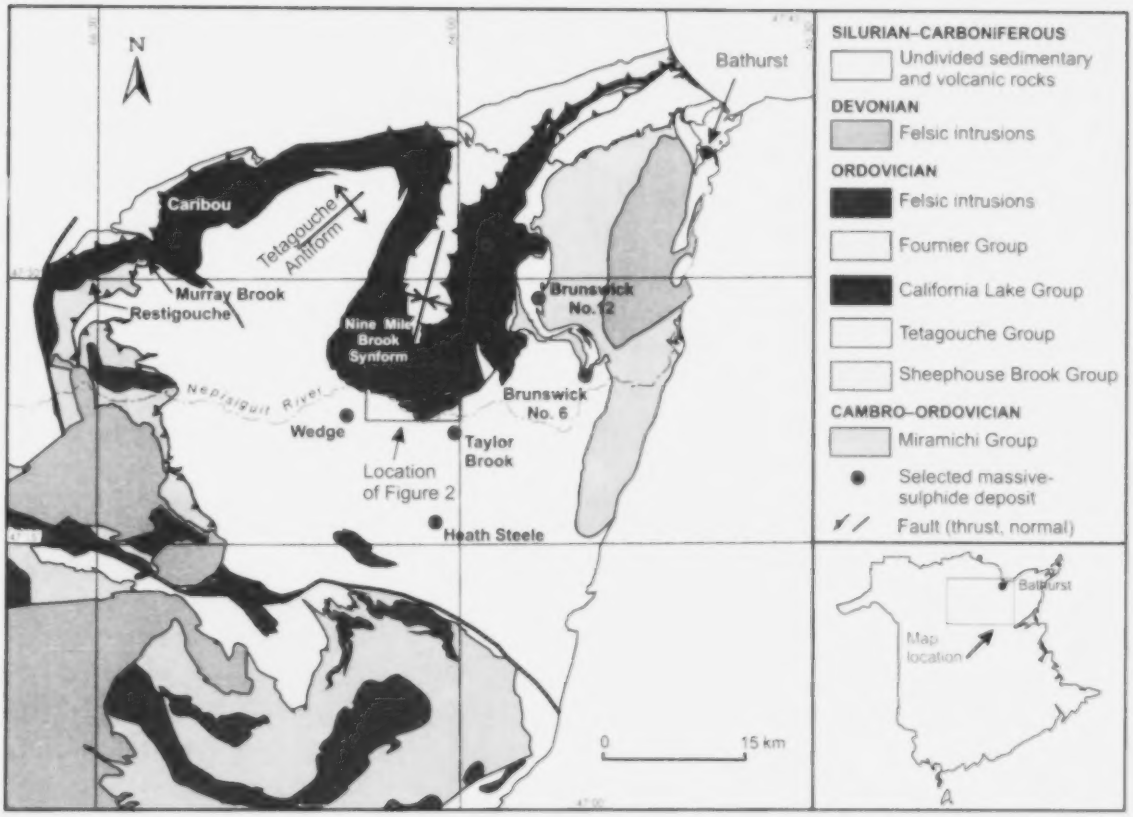

Figure 1. Major geological divisions of the Bathurst Mining Camp (after van Staal et al. 2003).

Nepisiguit A deposit contains an estimated resource of 1.54 million tonnes grading 2.8% Zn, 0.6% Pb, and 0.4% Cu; the Nepisiguit B deposit contains approximately 1.36 million tonnes grading 1.9% Zn, 0.4% Pb, and 0.1% Cu; and the Nepisiguit C deposit contains an estimated resource of 0.64 million tonnes grading 2.1% Zn, 0.7% Pb, and 0.4% Cu (Goodfellow and McCutcheon 2003; McCutcheon et al. 2003).

This paper documents the stratigraphic setting of the Nepisiguit A volcanogenic massive-sulphide deposit, describes the hydrothermal alteration associated with it, and attempts to determine its relationship to the Nepisiguit B and C deposits. The present work is part of ongoing mineral deposit studies in the BMC, which aim to reappraise the stratigraphic position of known sulphide deposits in the regional tectonostratigraphy, as well as to document the macroscopic and chemical characteristics of sulphide lenses and their associated hydrothermal alteration.

Previous Work

In 1954–1955 Kennco Exploration drilled three holes (M-1 to M-3) and conducted magnetometer surveys as well as vertical-loop electromagnetic (VLEM) surveys (Stevenson 1955). In 1956 the company conducted horizontal-loop electromagnetic magnetic (HLEM)

surveys and magnetometer surveys, together with stream and soil geochemical work. During that same year, Kennco drilled holes M-4 to M-18. Four of the holes (M-12, M-13, M-14, and M-17) intersected massive sulphides at the Nepisiguit A deposit (Stevenson and Brooks 1956), one (M-16) was drilled at the Nepisiguit B deposit, and eight (M-4 to M-11) were drilled at the Nepisiguit C deposit.

In 1957 Cominco conducted mapping and VLEM surveys and discovered gossan associated with the Nepisiguit B deposit. This was followed by gravity, HLEM, and diamond drilling (holes MO-1 and MO-2) work that intersected sulphides at the Nepisiguit A and B deposits (Cominco 1957). In 1957–1958, Cominco drilled six holes (MO-3 to MO-8) in the western ends of the Nepisiguit A and B deposits; the company also conducted mapping and Turam electromagnetic surveys that outlined the two deposits (Stevens 1958). Another 15 holes (NI-1 to NI-11, and NI-14 to NI-17) were concentrated in the eastern end of the Nepisiguit A deposit, and in the Nepisiguit B deposit. Results of this work showed that the B zone increases in grade with depth (Cominco 1958). Cominco later conducted a Turam electromagnetic survey (Bosschart 1959) and drilled four additional holes (MO-9, MO-10, MO-11, and NI-18; Cominco 1960, 1962).

Over the following several decades, work on the deposits included diamond drilling (holes NI-19 to NI-23; Cominco 1966); detailed geology, magnetometer, and gravity surveys (Cominco 1976, 1977); and a University of Toronto downhole electromagnetic (UTEM) survey on the Nepisiguit A and B deposits (Cominco 1982). Cominco also carried out line-cutting, magnetometer, and UTEM surveys in 1990, and drilled one more hole (NI-91-1) in 1991 (Roy 1996).

Noranda optioned the property in 1993 and conducted B-horizon soil geochemical surveys and magnetometer surveys as well as very-low-frequency electromagnetic, HLEM, and gravity surveys. It also undertook a program of geological mapping, diamond drilling (holes 94-1 to 94-4), trenching, and lithogeochemical analysis (Mitton 1994). Additional diamond drilling totalling six holes was reported by Roy (1996), and Perkins and Roy (1997).

Results of these foregoing investigations showed that the Nepisiguit A, B, and C deposits (Thomas et al. 2000) all display a distinct positive gravity anomaly and coincident conductivity and magnetic anomalies.

Regional Geology

The geology of the Bathurst Mining Camp (BMC) consists of a polydeformed Middle Ordovician sequence of bimodal volcanic and intercalated sedimentary rocks known collectively as the Bathurst Supergroup. The rocks were deposited in the Tetagouche–Exploits back-arc basin (van Staal and Fyffe 1991), which formed on an assortment of

Cambro–Ordovician siliciclastic sedimentary rocks (the underlying Miramichi Group) bordering the Gondwanan margin (van Staal and Fyffe 1991). During closure of the back-arc basin in the Late Ordovician to Early Silurian, rocks of the Bathurst Supergroup were dismembered and deformed in a subduction complex, resulting in convoluted tectonostratigraphic relationships.

The major stratigraphic divisions of the Bathurst Supergroup are the Sheephouse Brook, Tetagouche, California Lake, and Fournier groups (Fig. 1), each of which represents a major thrust nappe and contains numerous internal stratigraphic cut-outs and/or repetitions related to subsidiary thrust faults (van Staal et al. 2002, 2003). The Sheephouse Brook, Tetagouche, and California Lake groups all contain felsic and mafic volcanic and sedimentary rocks, whereas the Fournier Group is dominated by ocean-floor mafic volcanic and minor sedimentary rocks (McCutcheon et al. 1993; van Staal et al. 2002, 2003).

The Nepisiguit A, B, and C deposits are situated along the closure of the Nine Mile Brook Synform (Fig. 1), north of the faulted contact between the California Lake and Tetagouche nappes. The Tetagouche Group in this part of the BMC (Fig. 2) includes 1) the Nepisiquit Falls Formation that consists mainly of quartz-crystal tuffs, 2) the Flat Landing Brook Formation that comprises aphyric felsic volcanic rocks and minor basalts (Forty Mile Brook tholeiite), and 3) the Little River Formation that contains an interlayered sequence of basalts (Brunswick Mines alkali basalts) and sedimentary rocks (van Staal et al. 2002, 2003; Rogers and van Staal 2003).

The California Lake Group in the report area is divided into the Spruce Lake, Boucher Brook, and Canoe Landing Lake formations (Fig. 2). The Spruce Lake Formation is composed mainly of alkali feldspar-phyric felsic lavas and tuffs; a felsic unit containing both quartz and alkali-feldspar crystals; and tholeiitic basalt and sedimentary rocks, which constitute a minor component of the sequence. The Boucher Brook Formation conformably overlies the Spruce Lake Formation and comprises a sequence of sedimentary rocks interlayered with Camel Back Mountain alkali basalts (van Staal et al. 2002, 2003; Rogers and van Staal 2003). The Canoe Landing Lake Formation consists largely of alkali basalt but also includes tholeiitic basalts (the Nine Mile Brook tholeiite and Orvan Brook tholeiite). (The Orvan Brook tholeiite occurs in the vicinity of the Caribou deposit (Fig. 1) but is not present in the area of Figure 2.)

The footwall contact of the largely felsic sequence of the Spruce Lake Formation is a thrust fault that separates the lower part of the California Lake Group from the structurally underlying sedimentary and mafic volcanic sequence in the upper part of the Tetagouche Group (Fig. 2). Likewise, a thrust contact represents the boundary separating the dominantly mafic volcanic sequence of the Canoe Landing Lake Formation from the sedimentary and mafic volcanic sequence of the Boucher Brook Formation. Tholeiitic basalts of the Spruce

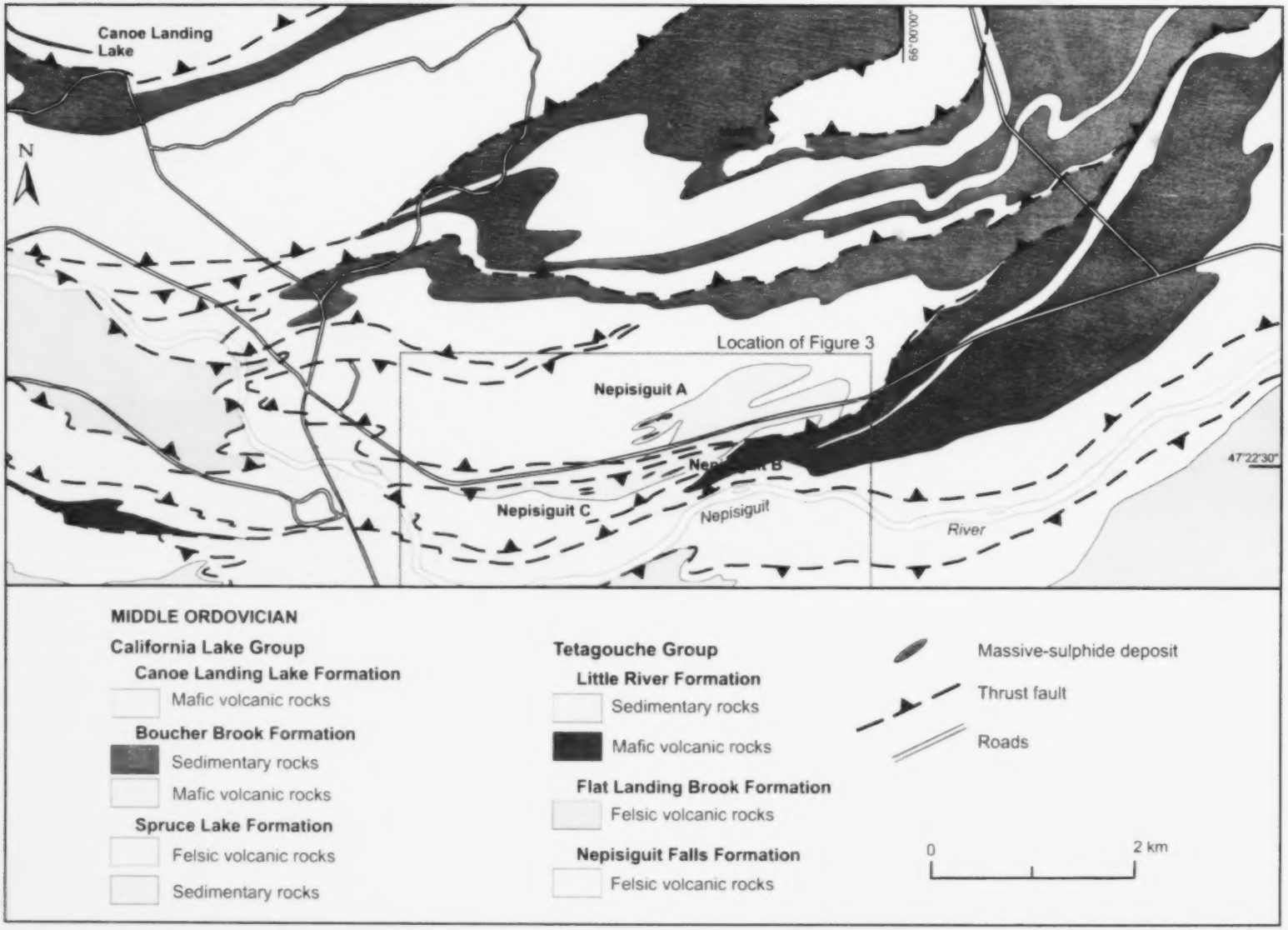

Figure 2. Regional geology in the central part of the Bathurst Mining Camp (modified after van Staal et al. 2002). See Figure 1 for map location.

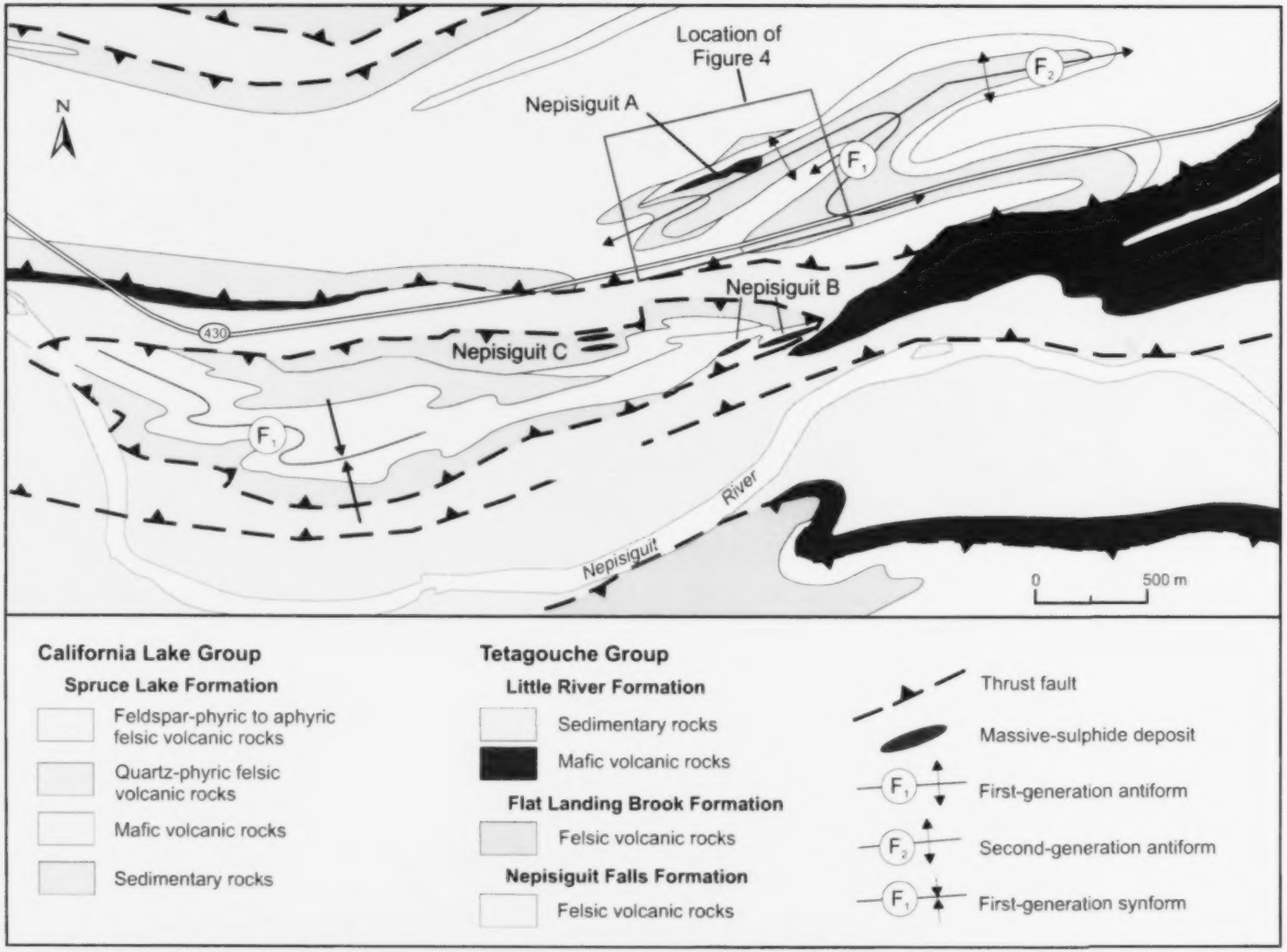

Figure 3. Geology in the vicinity of the Nepisiguit A, B, and C deposits (modified from Wilson 1993 and van Staal et al. 2002). Mafic rocks at the western end of the Nepisiguit A deposit have been projected to surface. See Figure 2 for map location.

Lake Formation and the Nine Mile Brook tholeiite of the Canoe Landing Lake Formation are interpreted to be coeval on the basis of their similar trace-element chemistry.

DEPOSIT GEOLOGY

Several regional mapping projects have been conducted in the immediate area of the Nepisiguit A, B, and C deposits (see Davies 1979; Wilson 1993; van Staal et al. 2002). According to van Staal et al. (2002), the thrust nappe hosting the Nepisiguit A deposit is composed solely of rocks from the California Lake Group (Fig. 3) and comprises felsic and minor mafic volcanic and sedimentary rocks, all assigned to the Spruce Lake Formation.

Geological interpretation in the area has been hampered by complex deformation features that include intense shearing at and near tectonic contacts, and polyphase folding. In addition, the relatively narrow unit widths and the juxtaposition of geophysically similar units (such as magnetic basalt, massive sulphides, and conductive graphitic sedimentary rocks) make it challenging to map solely on the basis of geophysics, as geophysical anomalies tend to overlap unit boundaries.

Although extensive diamond drilling has been conducted on the Nepisiguit A deposit (Stevens and Brooks 1956; McConnell 1958; Brooks 1960), most of the core from these drilling projects was not preserved. Fortunately, some drill core from the exploration program conducted by Noranda Inc. (Mitton 1994; Roy 1996; Perkins and Roy 1997) remains at the provincial core repository in Madran, northeastern New Brunswick. Four of these drill cores (94-1 through 94-4) were relogged and sampled for thin-sectioning, and samples from hole 94-2 were chemically analyzed (Fig. 4, 5).

A consistent stratigraphic younging direction was not evident in drill core and was not documented in industry drill logs (see Stevens and Brooks 1956; Stevens 1957; McConnell 1958; Brooks 1960; Mitton 1994; Roy 1996; Perkins and Roy 1997). However, relative positions of the massive sulphides and stringer sulphides indicate that younging is to the north (see *Massive-Sulphide Body*). Thus, the following stratigraphic descriptions of rocks of the Spruce Lake Formation at the Nepisiguit A deposit are based on an assumed north-younging sequence.

The lower part of the stratigraphic section consists of strongly deformed footwall sedimentary rocks dominated by black shale interlayered with thin (<10 cm) beds of fine-grained wacke (Fig. 5a, 5b, 6a). A thin (<4 m) carbonate-altered mafic volcanic unit occurs within the footwall sedimentary rocks at a depth of 265.2 m (856') in hole 94-2 (Fig. 4, 5a, 6b) in the western part of the Nepisiquit A deposit. A thicker section of footwall mafic volcanic rocks was intersected at depth in the eastern part of the deposit (Fig. 5b).

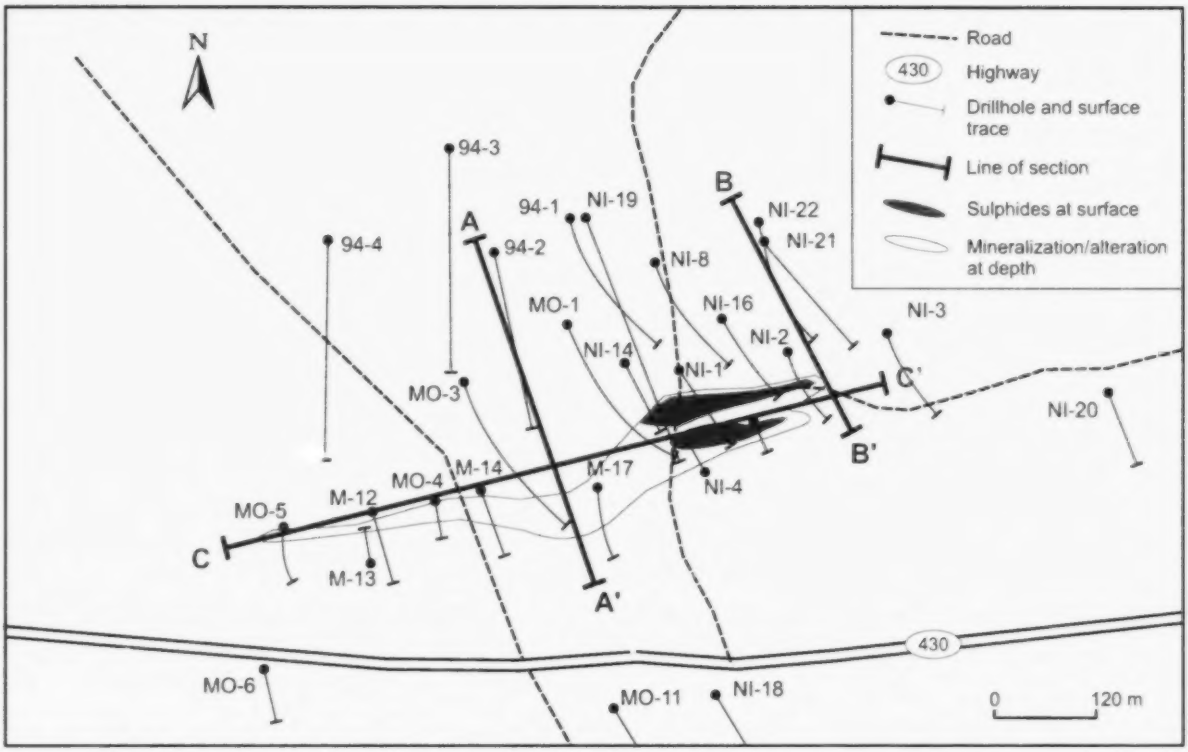

Figure 4. Drill plan of the Nepisiguit A deposit. Cross-sections A–A' and B–B' are shown on Figure 5, and longitudinal section C–C' is shown on Figure 16.

Mafic volcanic rocks are exposed at surface on the hanging wall of the sulphide lenses on the northern limb of a tight F, antiform (Fig. 3). Drill core from hole 94-2 indicates this mafic horizon consists of about 25 m of greyish green pillow basalt (Fig. 5a). The upper 6 m to 8 m of the basalt are brownish (Fig. 7a), a colour that may be the result of seafloor weathering. Epidote veins are locally abundant; other features include zoned quartz-carbonate veins and chains of small prismatic crystals tentatively identified as pyrophyllite (Fig. 7b). Pyrophyllite is an Al-rich alteration mineral commonly associated with acidic, high-sulphidation alteration of felsic rocks in epithermal systems, and its presence here is not easily explained.

Felsic volcanic rocks of the Spruce Lake Formation (Fig. 8a, 8b) in the hanging wall of the sulphide deposit range in texture from feldspar-phyric (crystal tuff) to aphyric (ash tuff). These volcanic rocks are interlayered with tuffaceous sedimentary rocks. The crystal tuffs are from 10 m to 16 m thick (Fig. 8a, 8b), and the tuffaceous sedimentary units are from 3 m to 11 m thick (Fig. 9a, 9b). Fragmental rocks are locally interlayered with the crystal tuff and tuffaceous sedimentary units. The fragmental rocks are heterolithic and contain matrix-supported, subrounded to angular clasts of black shale, wacke, feldspar-phyric rhyolite, and minor massive sulphides. Fragments range from less than 1 cm to greater than 8 cm in size (Fig. 9c, 9d). A thin interval of quartz-crystal-lithic tuff occurs near the bottom of the

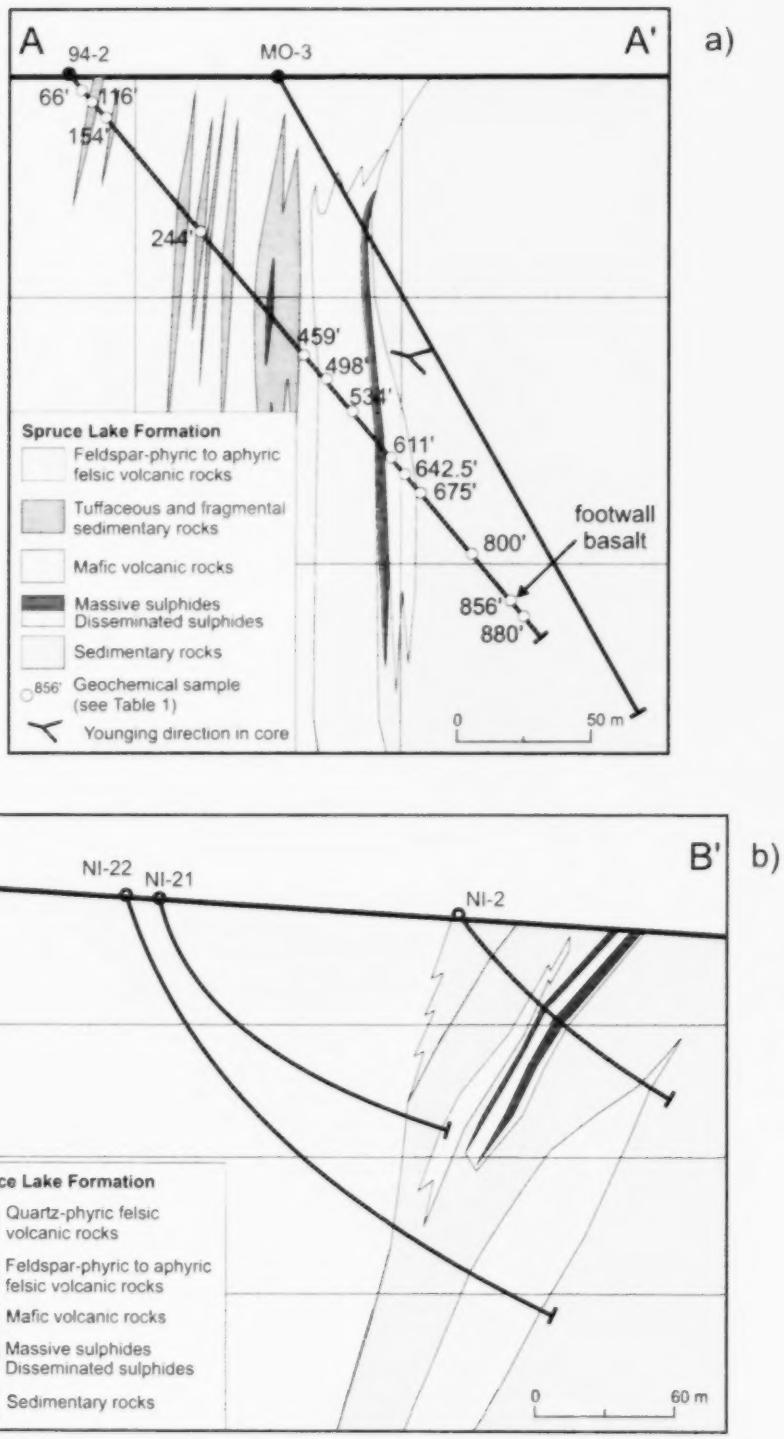

Figure 5. Geological cross-sections through the Nepisiguit A deposit. a) Cross-section A–A' from the western end of the deposit. b) Cross-section B–B' from the eastern end of the deposit (modified from Cominco 1966; Mitton 1994). See Figure 4 for section locations.

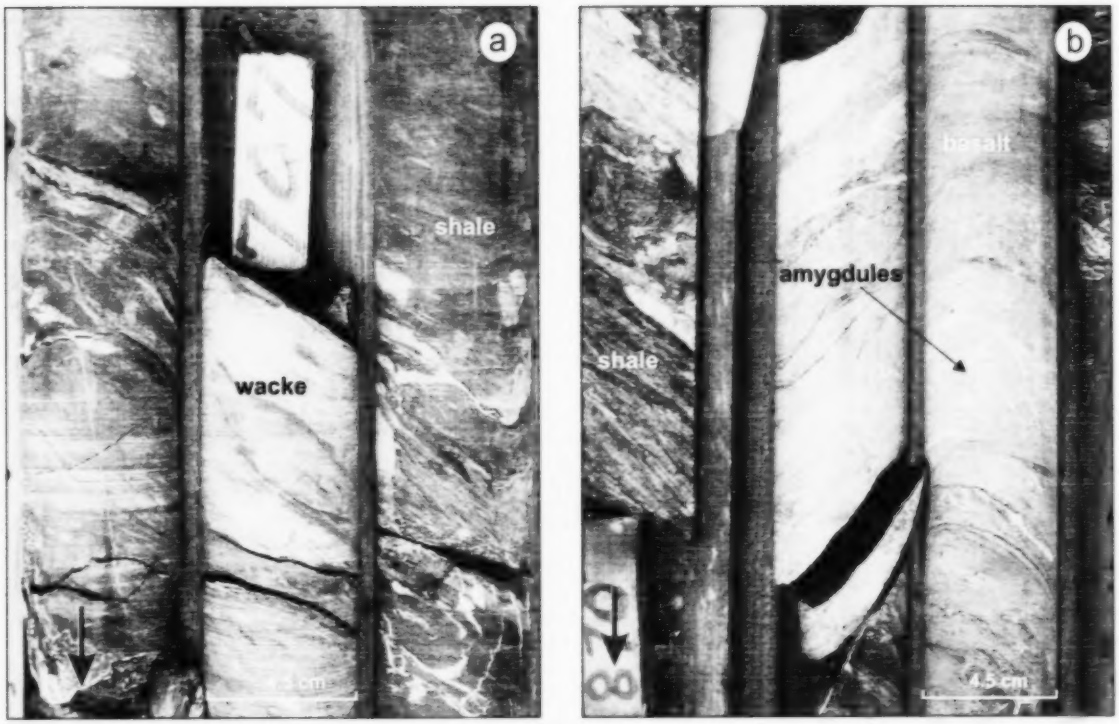

Figure 6. Sedimentary and mafic rocks in footwall of the Nepisiguit A deposit. a) Shale and wacke in DDH 94-2 at 215.5 m (707'). b) Mafic unit (light colour) and sedimentary rocks (light and dark mottled) in DDH 94-2 at 265.2 m (870'). Red arrows indicate the uphole direction.

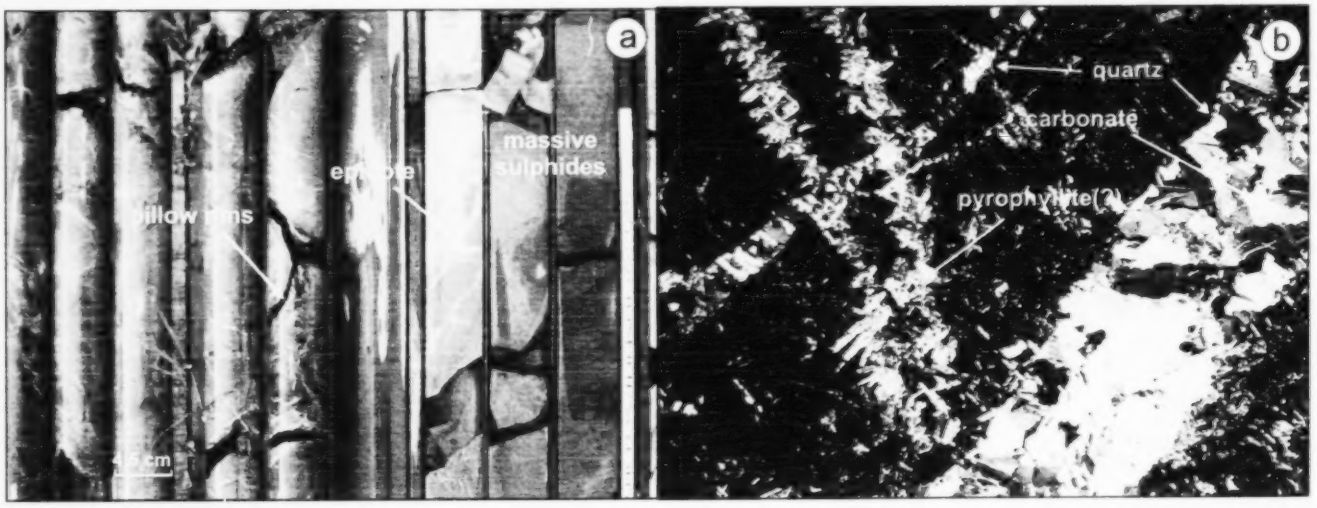

Figure 7. Mafic rocks in immediate hanging wall of the Nepisiguit A deposit. a) Pillowed mafic rocks immediately overlying massive sulphides from DDH 94-2 at 176.2 m (578'). Red arrow indicates the uphole direction. b) Photomicrograph of mafic rock in a). Shows altered groundmass, a zoned quartz-carbonate vein, and chains of crystals tentatively identified as pyrophyllite.

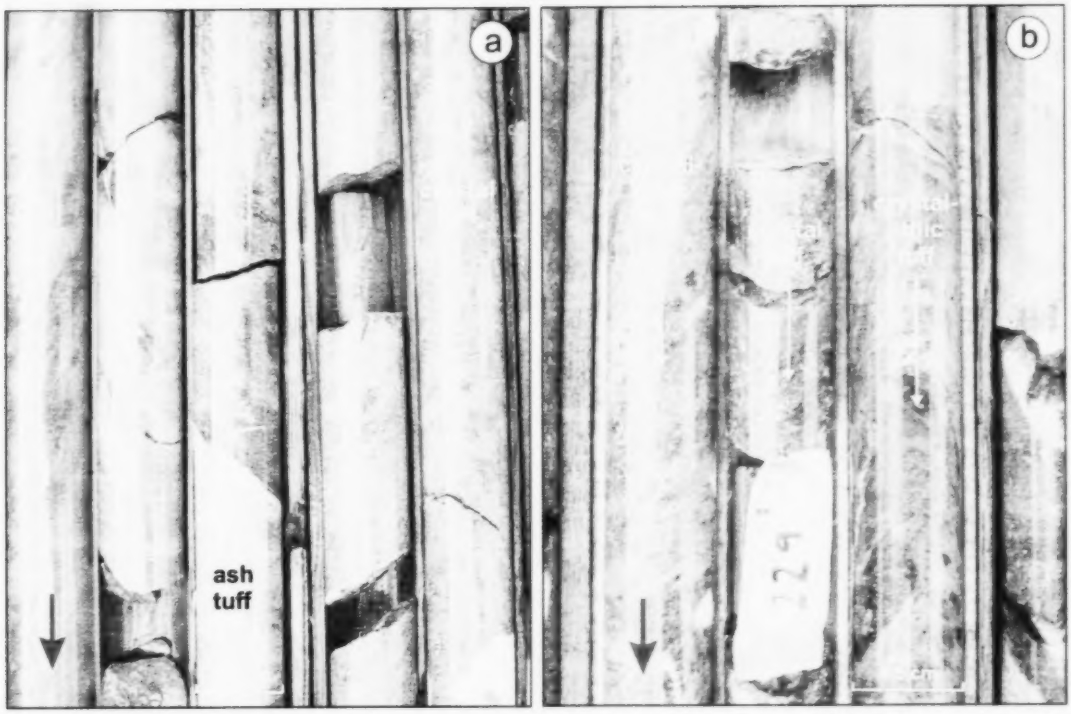

Figure 8. Felsic volcanic rocks in hanging wall of the Nepisiguit A deposit. a) Contact between ash tuff (light brown) and crystal tuff (grey) from DDH 94-2 at 92.7 m (304'). b) Crystal tuff and crystal-lithic tuff from DDH 94-2 at 70 m (229'). Red arrow indicates the uphole direction (and the stratigraphic tops.)

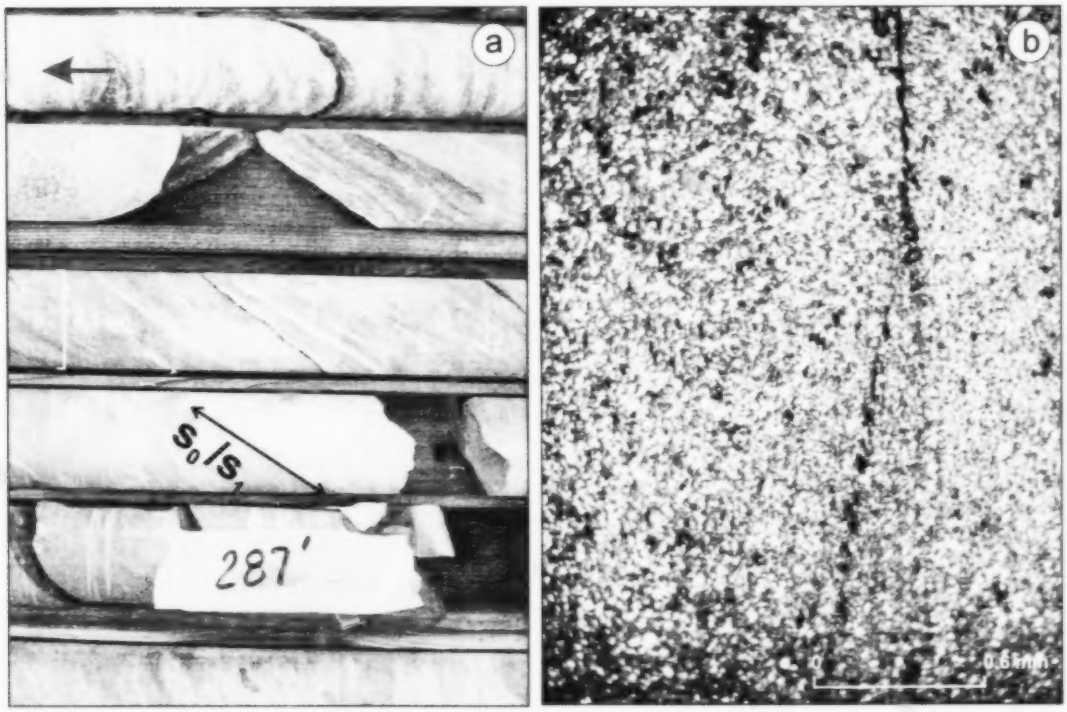

Figure 9. Sedimentary and fragmental rocks in the hanging wall of the Nepisiguit A deposit. a) Interlayered tuffaceous siltstone and crystal tuff from DDH 94-2 at 87 m (287'), showing composite bedding-cleavage fabric. Red arrow indicates the uphole direction. b) Photomicrograph of the sedimentary rock in a).

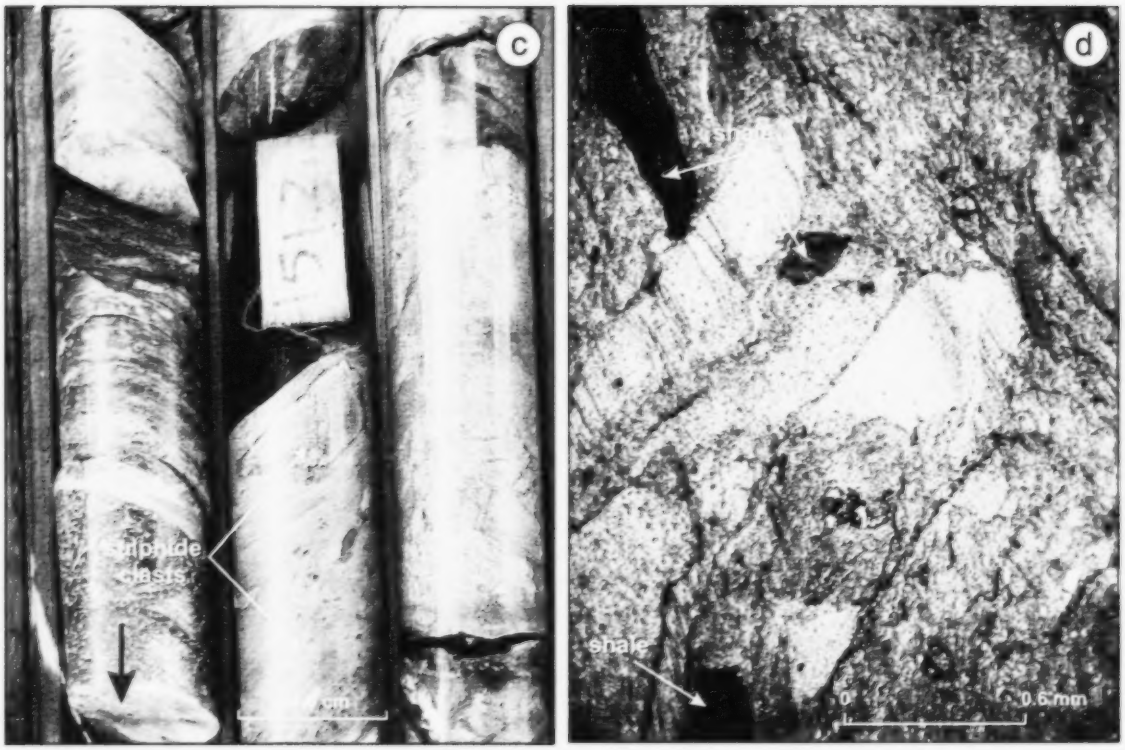

Figure 9 *(cont'd).* Sedimentary and fragmental rocks in the hanging wall of the Nepisiguit A deposit. c) Fragmental rock with clasts of crystal tuff, clastic sedimentary rocks, and massive pyrite from DDH 94-2 at 151.2 m (496'). Red arrow indicates the uphole direction. d) Photomicrograph of the fragmental rock shown in c).

hanging-wall sequence close to the contact with the Spruce Lake mafic flows in core 94-2 but is too narrow to show on Figure 5a. The same rock type has been reported (Cominco 1958) from a similar position in drillhole NI-2 (Fig. 5b).

The relationship between the Nepisiguit A deposit and the Nepisiguit B and C deposits is not well understood. Interpretation is hampered by the lack of surviving drill core from the Nepisiguit B and C deposits and by the apparently increasing intensity of deformation toward the south. Drill logs and sections from the Nepisiguit B and C deposits (see *Previous Work*) suggest that both deposits occur within a belt of locally graphitic sedimentary rocks. However, mafic rocks at the Nepisiguit B deposit occur in the immediate hanging wall and the footwall, whereas at the Nepisiguit C deposit they apparently occur only in the footwall.

LITHOGEOCHEMISTRY

Representative samples of the Nepisiguit A deposit were collected from drill core 94-2 for lithogeochemical analysis. They were assayed for major- and trace-element data by Activation Laboratories of Ancaster, Ontario, using a variety of analytical methods. These methods are summarized below.

- The major elements and Ba, Be, V, and Y were analyzed by inductively coupled plasma-optical emission spectrometry (ICP-OES), using lithium metaborate/ tetraborate fusion.
- Bi, Cs, Ga, Ge, Hf, In, Nb, Rb, Sn, Sr, Ta, Th, U, Zr, and the rare-earth elements were analyzed by mass spectrometry, using lithium metaborate/tetraborate fusion.
- Cd, Cu, Ni, Pb, and S were analyzed by ICP-OES, using total digestion.
- Ag, As, Au, Co, Cr, Hg, Mo, Sb, Sc, Se, W, and Zn were analyzed by instrumental neutron activation analysis.

Analytical data from these samples are listed in Table 1. These drill core samples and their analyses are hereafter referred to as the study samples and study data, respectively.

An additional set of lithogeochemical data from the deposit area was compiled from analyses conducted for Noranda Inc. by X-ray Assay Laboratories of Don Mills, Ontario, using a combination of x-ray fluorescence and ICP-OES methods for major- and trace-elements. Mitton (1994) and Roy (1996) describe the data, detection limits, and analytical methods. These combined Noranda samples and their analyses are hereafter referred to as the Noranda samples and Noranda data, respectively.

Both the current study data from core 94-2 and the older Noranda data have been plotted on the geochemical discrimination diagrams shown in Figures 10 to 14; these diagrams and their interpretations are discussed below.

Volcanic Rocks

The study data and Noranda data on the Na_2O+K_2O vs SiO_2 discrimination diagrams (Fig. 10a, 10b) indicate that SiO_2 values for the mafic samples lie below 47%, whereas those for the felsic volcanic samples are somewhat higher (55% to ≥75% SiO_2).

Although Na_2O+K_2O values in the felsic rocks are normal, the geochemical data (Table 1) show that K_2O values are unusually high and that Na_2O values are very low, commonly below detection limits. These alkali-element trends are typical of host volcanic rocks from the Bathurst Mining Camp. They are believed to be a function of seafloor alteration coupled with the effects of regional greenschist-grade metamorphism, which together resulted in high mobility of alkali elements (Lentz 1996). Consequently, high-field-strength elements are generally used to classify rocks in the Bathurst Mining Camp, because they are less mobile under the conditions described above.

On the Zr/TiO_2 vs Nb/Y diagram, felsic volcanic rocks fall into three populations (Fig. 10c, 10d):

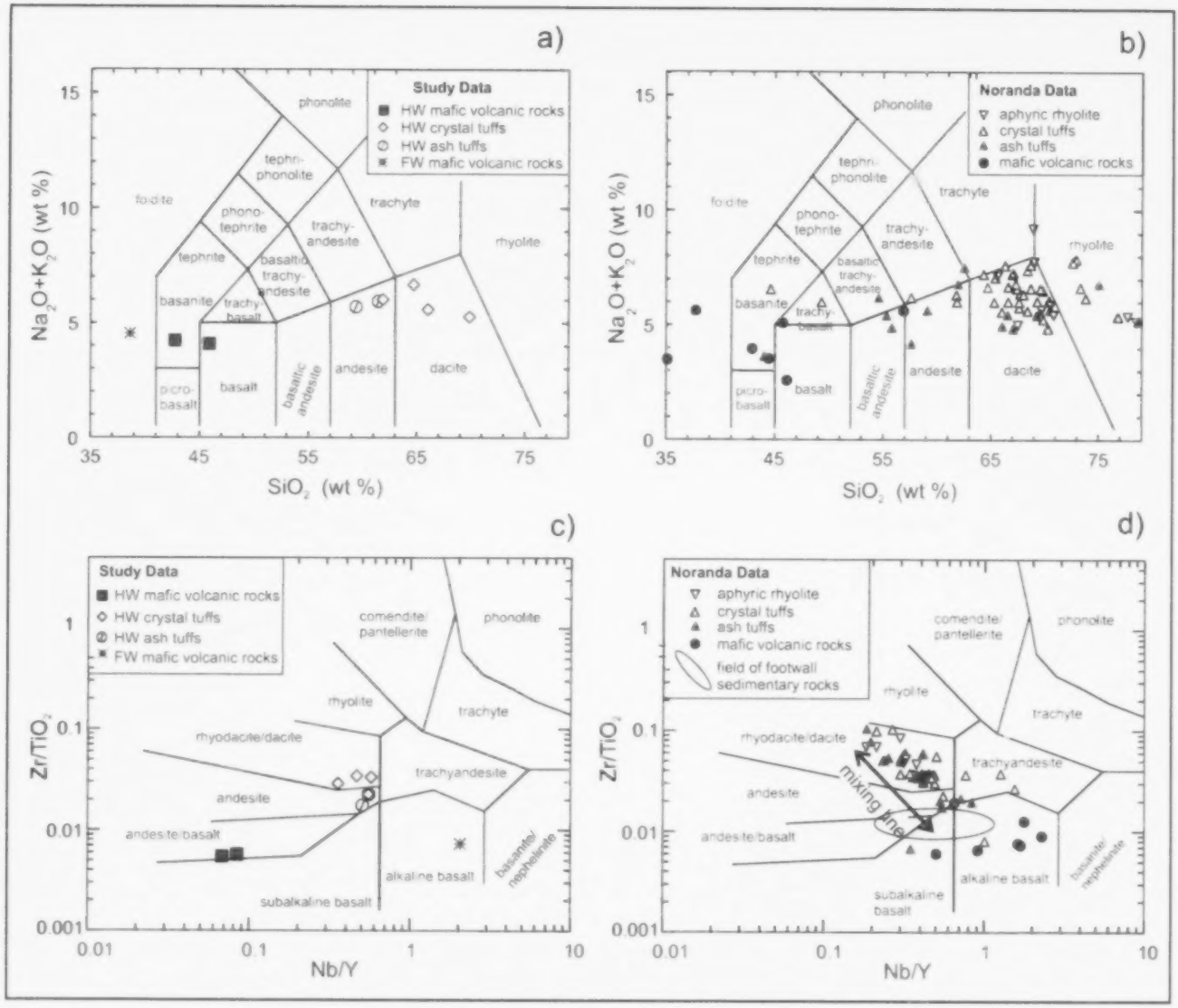

Figure 10. Major- and trace-element lithogeochemical discrimination diagrams for volcanic rocks from the Nepisiguit A deposit. a) and b) show the Na_2O+K_2O vs SiO_2 diagram (after Le Maitre et al. 1989) for the study data and Noranda data, respectively. b) and c) show the Zr/TiO_2 vs Nb/Y diagram (after Winchester and Floyd 1977) for the study data and Noranda data, respectively. Study data are from this report; Noranda data are from Mitton (1994) and Roy (1996).

- crystal tuffs from the study samples that cluster largely in the rhyodacite-dacite field (Zr/TiO_2 is typically about 0.035) and that have a strong negative correlation with the Nb/Y ratio,
- ash tuffs and tuffaceous sedimentary rocks from the study samples with Zr/TiO_2 values that are evenly distributed over a range of 0.015 to 0.10 and that display a strong negative correlation with Nb/Y, and
- a few samples of aphyric rhyolite collected by Noranda (Fig. 10d) that have Zr/TiO_2 values of between 0.03 and 0.10.

Table 1. Lithogeochemical data from host rocks at the Nepisiguit A sulphide deposit, and sulphide chemistry from the Nepisiguit A and B deposits, Bathurst Mining Camp.

Sample	94-2-0066	94-2-0116	94-2-0141.5	94-2-0154	94-2-0244	94-2-0459	94-2-0498	94-2-0534	94-2-0611	94-2-0642.5	94-2-0675	94-2-0800	94-2-0856	94-2-0880	Nepisiguit B	Nepisiguit A
Rock Type	crystal tuff	ash tuff	crystal tuff	crystal tuff	ash tuff	crystal tuff	HW mafic	HW mafic	altered zone	altered zone	shale	shale	FW mafic	shale	massive sulphides	massive sulphides
SiO_2	64.68	59.35	66.00	61.84	61.43	69.85	42.71	45.87	47.28	55.34	52.03	52.97	38.58	63.37	4.5	13.8
Al_2O_3	16.64	17.23	15.46	17.7	18.38	14.31	15.21	15.54	21.45	18.93	24.49	24.05	13.16	12.25	3.3	11
Fe_2O_3	4.95	8.4	6.06	6.53	7.01	4.66	8.65	10.48	12.86	9.35	8.24	8.23	9.54	7.36	55.9	43.4
MnO	0.14	0.21	0.11	0.20	0.08	0.10	0.49	0.30	0.02	0.12	0.14	0.12	0.32	0.05	0.09	0.06
MgO	2.28	2.5	2.56	1.82	1.58	1.82	8.85	7.61	0.52	0.62	1.98	2.08	3.81	3.17	0.49	0.29
CaO	0.29	0.31	0.25	0.15	0.16	0.37	6.51	6.76	0.18	0.38	0.19	0.18	12.51	2.34	0.1	0.15
Na_2O	3.72	1.72	2.96	1.83	0.29	3.18	3.71	4.07	0.29	0.26	0.82	0.86	3.49	1.75	0.03	0.09
K_2O	2.94	3.98	2.63	4.18	5.64	2.08	0.51	0.01	5.83	4.94	5.59	5.43	1.02	2	0.59	3.04
TiO_2	0.57	0.87	0.78	1.02	0.90	0.45	1.02	1.07	0.89	0.82	1.13	1.03	1.95	0.63	0.18	0.5
P_2O_5	0.19	0.21	0.16	0.1	0.12	0.15	0.09	0.09	0.005	0.005	0.1	0.1	0.44	0.08	0.03	0.07
LOI wt %	3.34	4.74	2.94	3.81	4.26	2.41	11.5	7.43	10.83	8.49	5.16	4.67	14.11	6.87	29.9	25.5
Total wt %	99.73	99.52	99.9	99.18	99.83	99.4	99.26	99.19	99.96	99.18	99.86	99.72	98.32	99.89	109.8	108
S	0.007	0.533	0.012	0.455	2.44	0.044	0.035	0.0005	11.8	8.72	1.06	0.543	0.425	1.12	43.7	35.3
Ag	0.25	0.25	0.25	0.25	0.6	0.25	0.25	0.25	0.9	24.9	0.25	0.25	0.6	0.25	16	4.2
As	12	19	13	9	10	12	64	6	167	200	20	12	13	13	400	540
Au	0.5	0.5	0.5	0.5	0.5	5	0.5	3	20	36	9	0.5	0.5	9	162	100
Ba	312	520	311	519	622	270	70	44	648	556	826	903	207	476	37	360
Be	3	4	3	4	5	2	2	1	2	3	5	5	1	2	< 0.05	1.4
Bi	0.1	0.3	0.05	0.2	0.2	0.2	0.1	0.05	0.05	0.05	0.05	0.05	0.05	0.05	26	30
Cd	0.25	0.25	0.25	0.25	1.4	0.25	1	1.4	3.6	15.3	0.25	0.7	0.5	0.5	56	7.2
Co	8.8	27.1	12.9	21.2	26.5	10.9	46.3	51.2	40.9	24.3	28.1	19.6	39.8	30.3	91	74
Cr	38.3	89.3	39	91.5	86.3	31.3	335	370	102	84.3	112	103	273	177	60	92
Cs	2.5	4.2	2.2	4.7	5.9	2.1	0.9	0.4	6.7	6.3	8.3	8.4	1.3	2.7	1.4	2.7
Cu	8	50	5	47	108	53	2	2	2030	70	35	30	48	62	1200	860
Ga	26	25	23	27	28	21	13	10	28	23	32	34	16	16	7	12
Ge	1.7	1.8	1.8	1.8	1.9	1.3	1.6	2.4	0.9	0.8	1.6	2.2	0.9	0.5		
Hf	5.9	4.8	6.8	7	6.1	4.9	1.8	1.7	3.8	3.4	4.8	4.1	3.8	3	1.2	2.5
Hg	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	5	0
In	0.05	0.05	0.05	0.05	0.2	0.05	0.4	0.1	0.5	0.2	0.05	0.05	0.05	0.05	9.4	0.65
Mo	1	1	1	1	1	1	1	1	1	1	1	1	1	10	2.5	2
Nb	17.2	17.2	17.1	19.9	20.5	14	1.6	1.5	19.8	17.2	23.6	23.3	41.2	8.1	4.4	10
Ni	17	37	22	36	49	13	133	140	38	43	60	49	102	158	15	29
Pb	2.5	13	2.5	2.5	29	9	2.5	2.5	632	> 5000	43	42	2.5	15	6700	380
Rb	132	175	117	200	252	104	26	< 2	240	206	226	219	32	66	23	120
Sb	1.3	3.3	1.3	3.5	5.5	2.2	3.5	5.4	40.3	37.6	7	3	0.8	9.4	94	98
Sc	14.3	18.9	16.4	19.8	18.1	11.8	36.6	41	22.7	15.9	22.8	22.6	22.7	18.6	2.9	8.8
Se	0.25	0.25	0.25	0.25	0.25	0.25	0.25	0.25	11.4	0.25	0.25	0.25	0.25	0.25	59	37
Sn	5	3	4	3	20	6	13	6	11	31	3	4	0.5	0.5	310	16
Sr	37	29	31	27	17	46	144	273	39	58	79	76	326	99	0	16
Ta	1.4	1.3	1.2	1.5	1.5	1.1	0.1	0.05	1.4	1.3	1.7	1.7	2.6	0.5	0.3	0.7
Th	18.3	13.3	19.5	15.4	16.2	15.3	0.16	0.12	16.9	15	19.4	20.7	3.03	5.9	3.3	8.5
U	6.26	3.04	4.86	3.78	4	5.19	0.12	0.1	2.39	2.89	4.2	3.54	0.93	6.03	2.3	2.6
V	61	124	85	137	117	43	193	213	125	101	135	127	191	154	20	68
W	0.5	4.0	0.5	2.0	2.0	2.0	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	4.0	0
Y	30	34	48	36	37	30	19	22	36	42	46	49	20	22	8.8	17
Zn	56	54	46	36	691	48	53	78	10000	9080	144	146	80	94	23000	7400
Zr	189	153	225	231	202	153	58	58	125	110	158	131	141	103	40	83

HW = hanging wall; FW = footwall. Nepisiquit A and B data are compiled from McClenaghan et al. 2003.
Major oxides in wt %. All trace elements in ppm except S (wt %), Au (ppb), and Hg (ppb).

Table 1 *(cont'd).* Lithogeochemical data from host rocks at the Nepisiguit A sulphide deposit, and sulphide chemistry from the Nepisiguit A and B deposits, Bathurst Mining Camp.

Sample	94-2-0066	94-2-0116	94-2-0141.5	94-2-0154	94-2-0244	94-2-0459	94-2-0498	94-2-0534	94-2-0611	94-2-0642.5	94-2-0675	94-2-0800	94-2-0856	94-2-0880	Nepisiguit B	Nepisiguit A
Rock Type	crystal tuff	ash tuff	crystal tuff	crystal tuff	ash tuff	crystal tuff	HW mafic	HW mafic	altered zone	altered zone	shale	shale	FW mafic	shale	massive sulphides	massive sulphides
La	54.7	45.6	59.7	45.8	51.9	42.1	2.91	2.52	57.4	42.8	60.6	64.6	24.7	22.3	13	29
Ce	117	106	127	97.8	112	88.9	8.7	8.1	115	89.2	123	131	50.4	45.7	26	62
Pr	13.9	11.4	15.2	11.5	12.7	10.7	1.41	1.36	14.2	10.5	14.5	15.5	6.06	5.44	2.9	6.9
Nd	49.1	40.3	56	41	45	38.2	7.25	7.38	50	36.7	51	54.1	22.5	19.7	11	26
Sm	10.1	8.25	11.6	8.11	8.97	7.88	2.23	2.33	10.2	7.75	10.1	10.8	4.9	4.12	2.1	4.8
Eu	1.45	1.67	2.05	1.57	1.99	1.18	1.0	0.644	2.5	1.79	2.12	2	1.54	0.912	1	1.5
Gd	8.35	7.01	10.9	7.15	7.86	6.77	2.97	3.13	8.65	7.46	9.22	10	4.61	4.05	1.8	4
Tb	1.2	1.17	1.73	1.17	1.25	1.05	0.54	0.61	1.26	1.25	1.44	1.57	0.71	0.67	0.26	0.62
Dy	6.03	6.03	9.26	6.35	6.61	5.49	3.35	3.72	6.3	6.96	7.76	8.41	3.75	3.74	1.5	3.5
Ho	1.08	1.1	1.69	1.23	1.26	1.03	0.69	0.76	1.18	1.34	1.49	1.62	0.73	0.77	0.29	0.62
Er	3.02	3.32	5.01	3.84	3.78	3.01	2.0	2.27	3.6	4.15	4.5	4.86	2.19	2.34	0.81	1.7
Tl	0.95	1.03	0.79	1.28	1.47	1.19	0.41	0.13	0.53	0.84	1.38	1.45	0.43	0.44	2.8	3.8
Tm	0.445	0.507	0.738	0.618	0.567	0.445	0.289	0.341	0.542	0.628	0.675	0.726	0.318	0.367	0.13	0.28
Yb	2.75	3.13	4.54	3.87	3.64	2.7	1.81	2.1	3.55	3.81	4.23	4.51	2.02	2.35	0.83	1.8
Lu	0.398	0.443	0.63	0.56	0.522	0.367	0.27	0.292	0.512	0.523	0.599	0.631	0.28	0.341	0.13	0.28
Tot REE	270.47	236.96	306.84	231.85	259.52	211.01	35.83	35.69	275.42	215.70	292.61	311.78	125.14	113.24	64.55	146.80
La_n/Lu_n	14.80	11.09	10.21	8.81	10.71	12.36	1.16	0.93	12.08	8.81	10.90	11.03	9.50	7.04	10.77	11.16
Eu*	17.85	14.78	21.86	14.80	16.32	14.20	5.00	5.25	18.26	14.78	18.76	20.20	9.24	7.94	3.78	8.52
Eu_n/Eu^*	0.48	0.67	0.56	0.63	0.73	0.49	1.19	0.73	0.81	0.72	0.67	0.59	0.99	0.68	1.57	1.05
Zr/TiO_2	0.06	0.03	0.05	0.04	0.04	0.06	0.01	0.01	0.02	0.02	0.02	0.02	0.01	0.03	0.04	0.03

HW = hanging wall; FW = footwall. All rare-earth elements in ppm. Nepisiquit A and B data are compiled from McClenaghan et al. 2003.

The ash tuffs overlap the range of Zr/TiO_2 and Nb/Y values exhibited by both the hanging-wall crystal tuffs and the footwall sedimentary rocks (Fig 10c, 10d). This spread in chemical signature may be a function of mixed sedimentary/felsic volcanic input.

Two samples of hanging-wall mafic rocks collected during this study (94-2-0498 and 94-2-00534; Fig. 5a) fall on the boundary of the subalkaline basalt and the andesite/basalt fields on the Zr/TiO_2 diagram (Fig 10c). Loss-on-ignition for these two samples is high (7.4% and 11.5%), indicating the possibility of hydrothermal alteration. In these samples, SiO_2 values (42.71% and 45.87%) are lower and MgO values are higher (7.61% and 8.85%) than the average Spruce Lake tholeiitic basalt.

A study sample of mafic rock from the footwall (Sample 94-2-0856; Fig. 5a, 10c) plots well within the alkaline basalt field (Nb/Y >2). Similarly, analyses of mafic volcanic rocks from the Noranda samples (Fig. 10d) plot mostly in the alkaline basalt field (Nb/Y >0.65) and give Zr/TiO_2 values that are equal to or slightly greater than that of the study sample.

Although most felsic samples fall within the calc-alkaline field (Fig. 11), mafic volcanic rocks from both the study and Noranda data sets tend to lie along an iron-enrichment trend. The two hanging-wall mafic samples collected during this study plot within the field of the Spruce Lake tholeiitic basalt on the Cr vs Nb/Y discrimination diagram. In contrast, the footwall alkali mafic sample falls in the field of the Orvan Brook tholeiite (Fig. 12a).

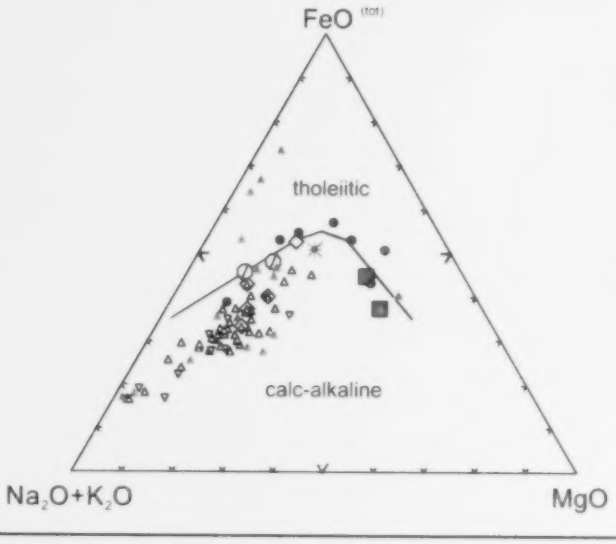

Figure 11. Na_2O+K_2O–$FeO^{(tot)}$–MgO diagram for volcanic rocks of the Nepisiguit A deposit (after Irvine and Baragar 1971).

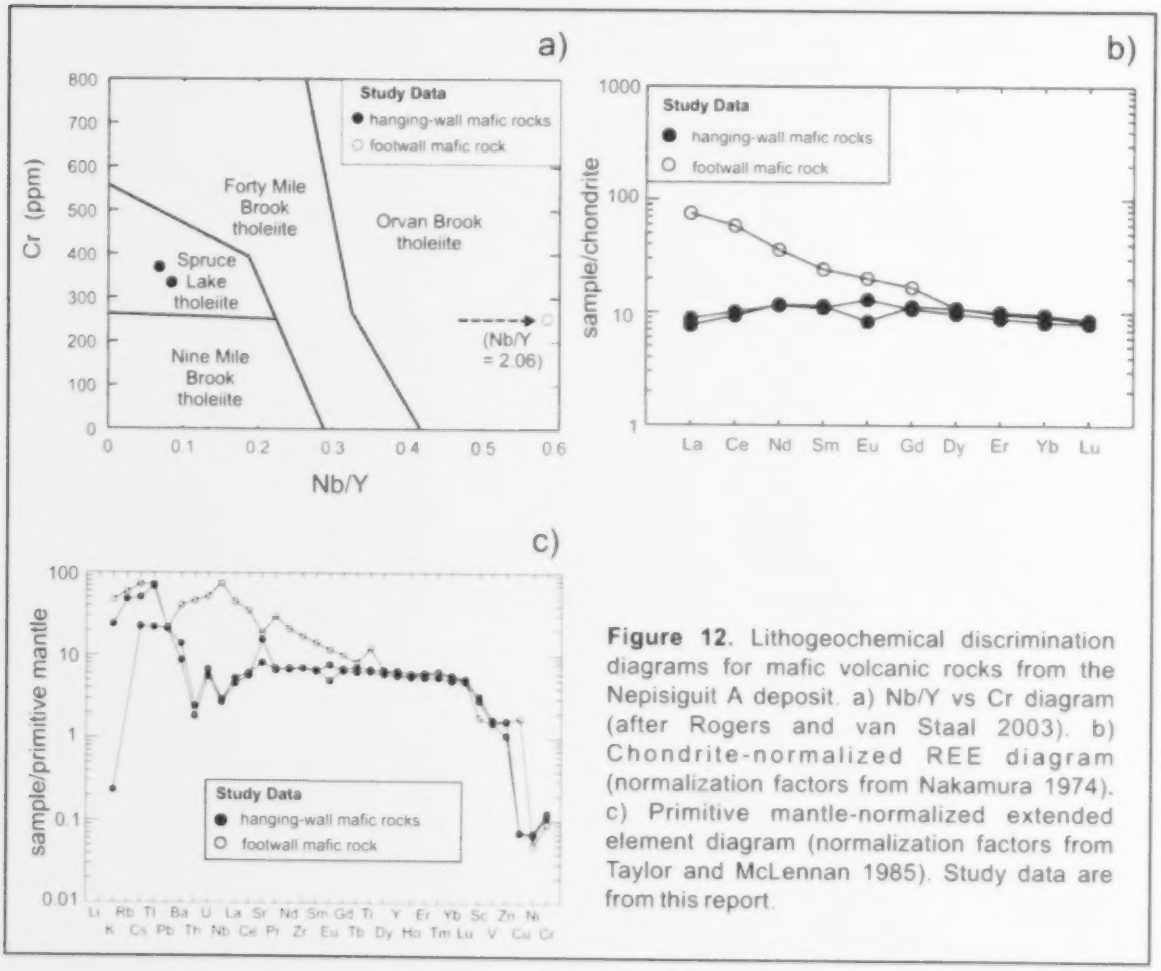
a)
Cr (ppm)
Nb/Y
Study Data
hanging-wall mafic rocks
footwall mafic rock
Forty Mile Brook tholeiite
Orvan Brook tholeiite
Spruce Lake tholeiite
Nine Mile Brook tholeiite
(Nb/Y = 2.06)
b)
sample/chondrite
La Ce Nd Sm Eu Gd Dy Er Yb Lu
c)
sample/primitive mantle
Li Rb Tl Ba U La Sr Nd Sm Gd Ti Y Er Yb Sc Zn Ni
K Cs Pb Th Nb Ce Pr Zr Eu Tb Dy Ho Tm Lu V Cu Cr

Figure 12. Lithogeochemical discrimination diagrams for mafic volcanic rocks from the Nepisiguit A deposit. a) Nb/Y vs Cr diagram (after Rogers and van Staal 2003). b) Chondrite-normalized REE diagram (normalization factors from Nakamura 1974). c) Primitive mantle-normalized extended element diagram (normalization factors from Taylor and McLennan 1985). Study data are from this report.

Trace-element values for the hanging-wall mafic samples are only somewhat similar to those for the average Spruce Lake tholeiitic basalt (Rogers and van Staal 2003). Specifically, Cr values are comparable, V values are 33% lower, and Ni values are more than twice those of the typical Spruce Lake basalt (Rogers and van Staal 2003). Cu values of the samples are extremely low (2 ppm), as compared with an average of 45 ppm Cu for the typical Spruce Lake basalt (Rogers and van Staal 2003). The average values for U (1.1 ppm) and Th (0.14 ppm) are 50% and 20%, respectively, of average values for the Spruce Lake basalt. The low Zn values (53–78 ppm) of these samples are more akin to those for felsic rocks than for mafic rocks.

Chondrite-normalized rare-earth-element (REE) patterns for the hanging-wall mafic rocks are fairly flat: light REE and heavy REE values are low, and middle REE (Nd–Dy) values are slightly higher (Fig. 12b). The average ΣREE for the two samples is approximately 35 ppm, the La_n/Lu_n values are 0.93 and 1.16, and the Eu_n/Eu^* values are 1.19 and 0.73. The REE profiles for these samples are very similar to those of other Spruce Lake tholeiitic basalts, although their absolute values are noticeably lower than the average reported by Rogers and van Staal (2003).

The chondrite-normalized REE profile for the footwall mafic sample is significantly enriched in light REEs, in comparison with the same profile for the hanging-wall samples (Fig 12b). The ΣREE for the footwall sample is approximately 125 ppm, the La_n/Lu_n value is approximately 9.5, and the Eu_n/Eu^* value is 0.99 (no Eu anomaly is present). The REE profile for this sample is similar to those of the Brunswick Mines, Canoe Landing Lake, and Camel Back Mountain alkali basalts (Rogers and van Staal 2003). Its absolute REE values are closer to those of the Canoe Landing Lake alkali basalt, although the latter rock is apparently less enriched in light REEs.

The extended primitive mantle-normalized profiles for hanging-wall and footwall mafic rocks likewise are noticeably different. Relative to the hanging-wall mafic rocks, the footwall sample is enriched by a factor of 10 in most large-ion lithophile elements (Ba, Th, U, Nb, La, and Ce), and by a factor of 2 to 4 in the elements K, Rb, and Cs (Fig. 12c). The higher field-strength elements (Sr–Ti) are also more enriched in the footwall mafic rock, but by a factor of less than 10.

Sedimentary Rocks

Major- and trace-element lithogeochemistry of sedimentary rocks in the Bathurst Mining Camp can be used to distinguish rocks deposited before or during felsic volcanism (that is, the Miramichi Group or lower parts of the Tetagouche and California Lake groups) from those deposited synchronous with or after mafic volcanism (that is, upper parts of the

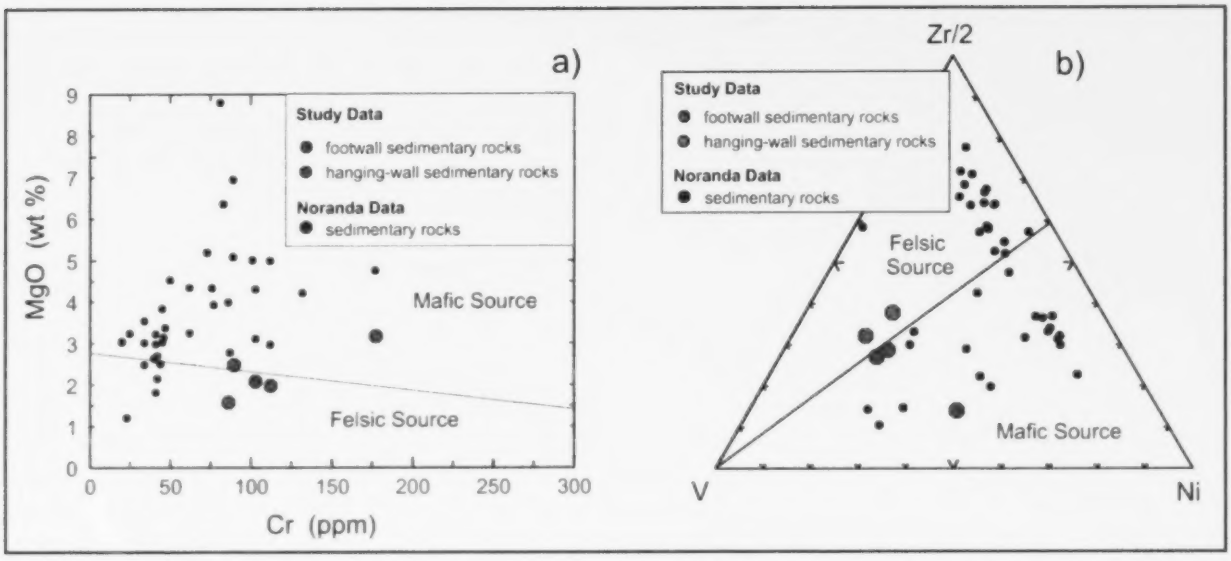

Figure 13. Lithogeochemical discrimination diagrams for sedimentary rocks from the Nepisiguit A deposit. a) Cr vs MgO diagram (after Rogers et al. 2003). b) V–Zr/2–Ni ternary diagram (after McCutcheon, unpublished data).

Tetagouche and California Lake groups; Rogers et al. 2003). These two rock series are designated as the 'Felsic Source' field and 'Mafic Source' field on Figure 13.

On the Cr vs MgO diagram (Fig. 13a), one sample of the footwall sedimentary rock plots well into the Mafic Source field; the two other samples plot in the Felsic Source field but are close to the field boundary. One sample of the hanging-wall sedimentary rock plots well into the Felsic Source field; the other sample lies in the Mafic Source field but is also adjacent to the boundary. Although data are limited, the proximity of the footwall and hanging-wall samples indicates derivation from similar sources. The hanging-wall samples appear to have a higher component of felsic input, which is consistent with their association with the Spruce Lake felsic volcanic sequence. Most Noranda samples fall within the Mafic Source field, suggesting that at least some of the high MgO values in the Noranda data reflect hydrothermal alteration rather than provenance.

As might be expected, the two hanging-wall samples (study samples) plot within the Felsic Source field on the V–Zr/2–Ni diagram (Fig. 13b), whereas the three footwall sedimentary samples (also study samples) fall within the Mafic Source field. The Noranda samples plot across both fields in this diagram, which, again, may be a function of alteration rather than provenance.

The importance of sedimentary rocks in terms of their control on volcanic massive-sulphide accumulation is well documented in the Bathurst Mining Camp (Goodfellow and McCutcheon 2003). Sedimentary rocks in the footwall sequence are significant because 1) they indicate the development of basins, which are important for collecting sufficient volumes of metalliferous brine to form large deposits, and 2) the accumulation of fine-grained pelagic sediments indicates a quiescent period lengthy enough to allow for deposit formation.

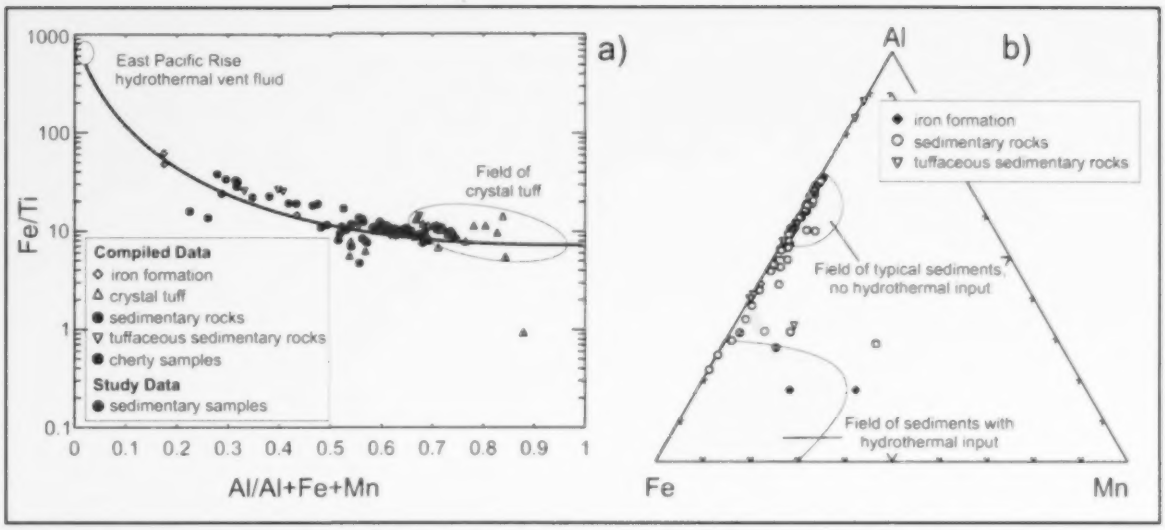

Figure 14. Lithogeochemical discrimination diagrams for the determination of hydrothermal input in some sedimentary rocks from the Bathurst Mining Camp (after Boström 1973). a) Fe/Ti vs Al /Al+Fe+Mn. b) Fe–Al–Mn ternary plot diagram. Compiled data are from Mitton (1994), Roy (1996), Perkins and Roy (1997), and this study.

Sedimentary rocks situated along strike from massive-sulphide deposits commonly show evidence of hydrothermal input that is directly attributable to venting fluids (Peter and Goodfellow 1996). In addition, sedimentary rocks stratigraphically positioned in the hanging wall of a sulphide-bearing horizon may show evidence of hydrothermal input that was imparted during waning stages of the hydrothermal system. Thus, determining the degree of hydrothermal input into a sedimentary rock can be valuable when assessing the sulphide-bearing potential of a particular horizon. This input can be measured using ratios that compare hydrothermally sourced elements (Fe+Mn) with detritally sourced elements (Al+Ti), as depicted in the Fe/Ti vs Al/Al+Fe+Mn and the Fe–Al–Mn diagrams (Fig. 14a, 14b).

The Fe/Ti vs Al/Al+Fe+Mn diagram (Fig. 14a) indicates that most sedimentary samples from the Nepisiguit A area have only a minor hydrothermal component. However, Noranda Inc. reported Algoma-type, banded cherty iron formation in drill core from south of the Nepisiguit B deposit, close to where the Little River Formation of the Tetagouche Group is known, on surface, to host iron formation but not massive-sulphide accumulations (Roy 1996). These samples of iron formation and chert provide evidence of hydrothermal input. Although assigned to the Little River Formation, these samples may be part of the Spruce Lake Formation, as the contact between sedimentary rocks of the Tetagouche Group and those of the California Lake Group has not yet been clearly defined in this area.

As expected, most of the sedimentary samples show a minimal hydrothermal component in the Fe–Al–Mn ternary plot (Fig. 14b). Sedimentary rocks from the hanging wall in drillhole 94-2 (study samples) show no evidence of hydrothermal input. One Noranda sedimentary sample from near the bottom of drillhole 94-1 has a relatively high hydrothermal component and is interpreted to lie close to the exhalite horizon. Other samples with elevated hydrothermal signatures were collected from trenches immediately along strike from the Nepisiguit B and C deposits (Mitton 1994; Roy 1996).

MASSIVE-SULPHIDE BODY

The Nepisiguit A deposit consists of two subparallel, easterly thickening, massive-sulphide lenses that strike approximately east-northeast and dip about 80°N (Fig. 4, 5). The southern lens was not intersected by any drill cores examined during this study and will not be discussed further. The larger, northern lens (Fig. 15, 16) has a strike length of approximately 380 m and extends downdip for about 280 m, where intersections of 8.93% Zn over 9.75 m have been reported (McAllister 1960; Mitton 1994). Sulphide intersections in the western end of the deposit are between 0.3 m and 3 m in thickness. The Nepisiguit A deposit as a whole contains an inferred resource of 1.54 million tonnes grading 2.8% Zn, 0.6% Pb, and 0.4% Cu (McCutcheon et al. 2003). Massive sulphides in the deposit are dominated by pyrite but include subordinate pyrrhotite, sphalerite, galena, and chalcopyrite.

In drill core 94-2 in the western part of the Nepisiguit A deposit (Fig. 4, 5a), massive sulphides are immediately overlain by basalt (Fig. 15a) and are underlain by a roughly 20 m thick sericite- or chlorite-altered zone of tuffaceous sedimentary rocks cut by quartz-sulphide veins (Fig. 15b). This altered, mineralized zone is interpreted to represent a feeder/stockwork zone, a feature that commonly underlies massive sulphides in volcanic massive-sulphide systems (Franklin et al. 1981). Drill logs and sections from assessment

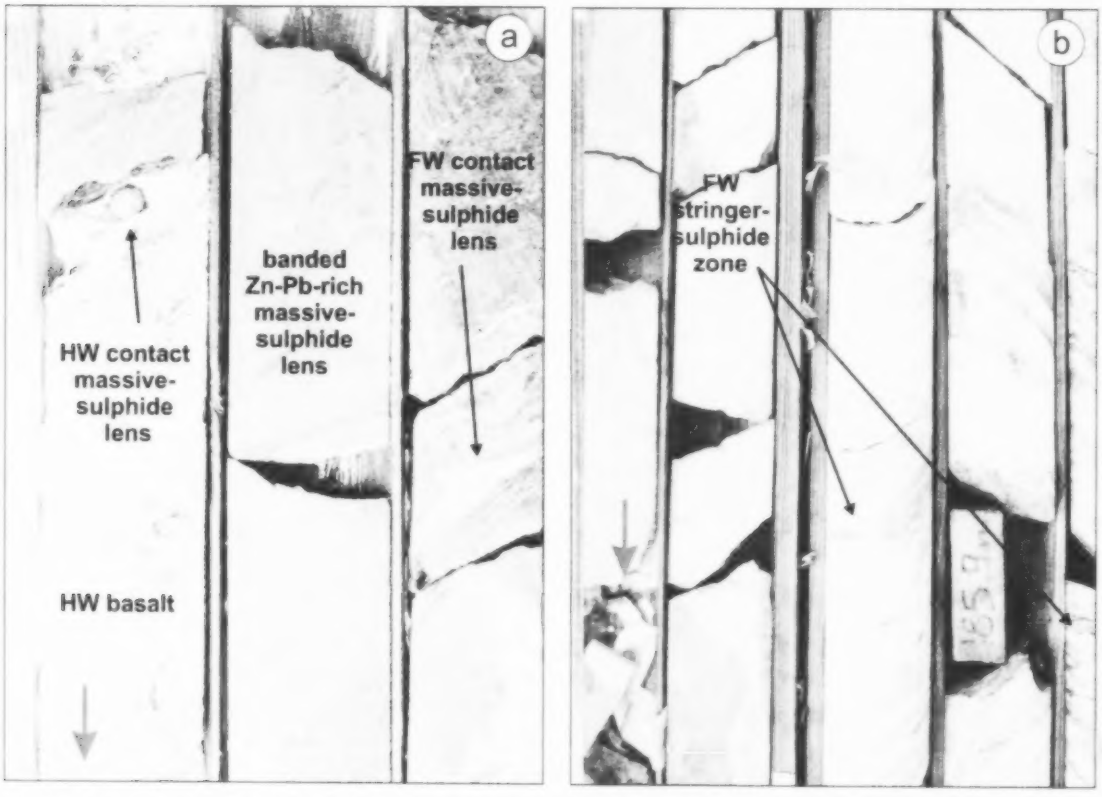

Figure 15. Massive- and stringer-sulphide mineralization from the Nepisiguit A deposit. a) Top of the massive-sulphide lens in contact with basalt in the hanging wall from DDH 94-2 at 176 m (577'). b) Stringer-sulphide zone mineralization beneath the massive-sulphide lens from DDH 94-2 at 185.9 m (611'). Red arrow indicates the uphole direction.

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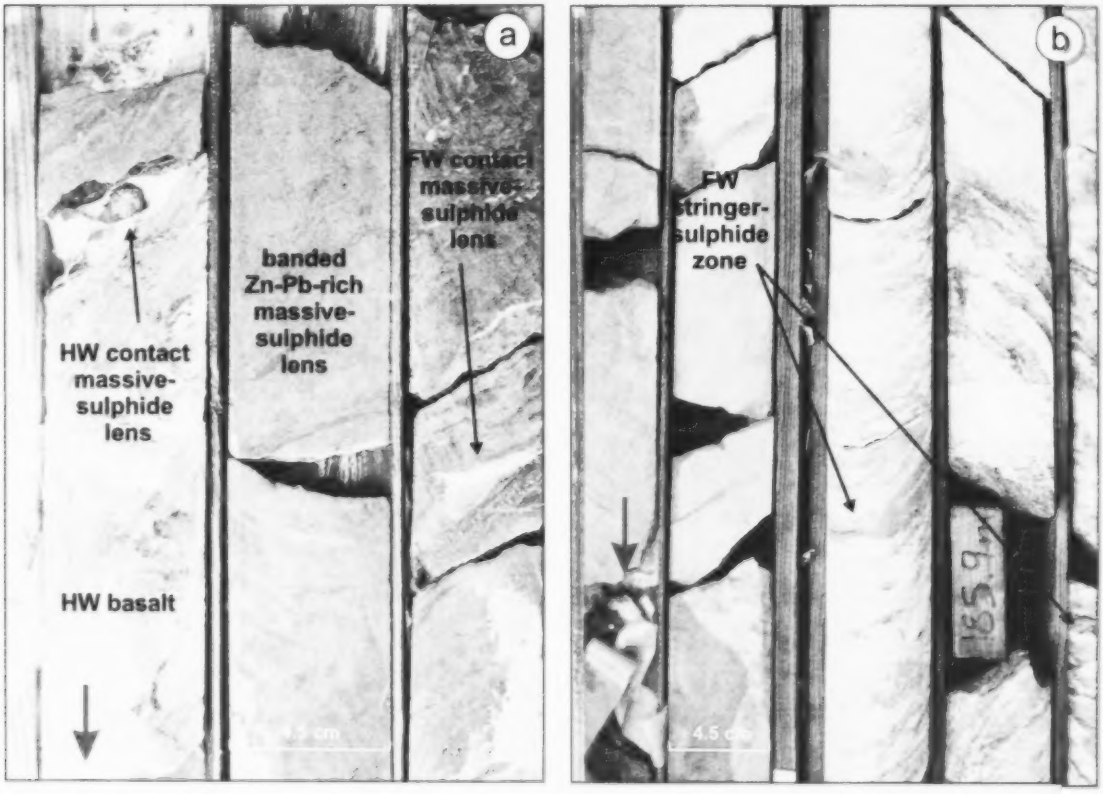

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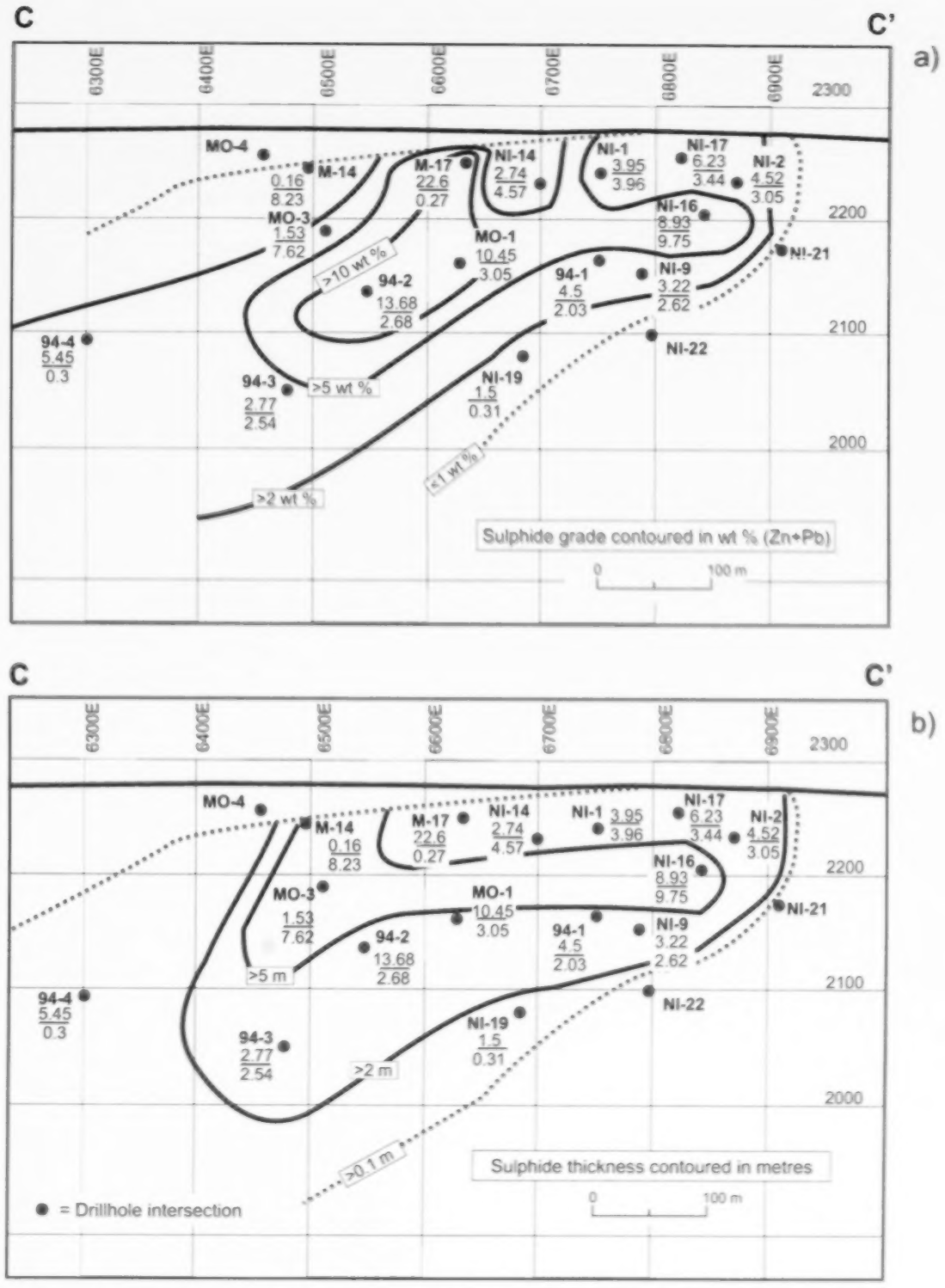

Figure 16. Longitudinal section C–C' through the northern lens of the Nepisiguit A deposit. Numbers indicate the value of Zn+Pb (in wt %)/width of assay interval (in metres) for each drillhole intersection (modified after Mitton 1994). a) The sulphide grade, contoured in wt % Zn+Pb. b) The sulphide thickness, contoured in metres. See Figure 4 for the line of longitudinal section.

reports suggest a similar relationship in the eastern part of the deposit, where massive sulphides reportedly are underlain by zones of disseminated and/or stringer sulphides (Fig. 5b).

Contoured Pb+Zn values (Fig. 16a) and contoured sulphide thicknesses (Fig. 16b) in a longitudinal section of the northern lens illustrate the westerly plunge of the Nepisiguit A deposit. The compiled assay data have been plotted on a Pb–Cu–Zn ternary diagram (Fig. 17). As with many Bathurst Mining Camp deposits, most sulphide data points from the deposit cluster along the Pb–Zn axis and have a Zn/Pb value of between 2 and 3. The average Zn/Pb value for samples containing higher than 0.01% Pb is approximately 12.7.

Sulphide Chemistry

Table 1 includes lithogeochemical analytical data for one sulphide sample from each of the Nepisiguit A and B deposits (McClenaghan et al. 2003). As with most Bathurst Mining Camp (BMC) deposits, massive sulphides in the two Nepisiguit deposits are characterized by elevated values of As, Au, Cd, Hg, In, Mo, Sn, and W in addition to Fe, Zn, Pb, and Cu (Table 1). The high concentrations of these elements, in particular Sn, indicate magmatic fluid input to the hydrothermal cell. The Sn content of BMC sulphide deposits also seems to increase with total sulphide tonnage. For instance, the average Sn content of the Brunswick No. 12 deposit (about 300 million tones) is 400 ppm, whereas the average Sn content for the smaller deposits (0.1 to 1 million tonnes) is substantially less (≥100 ppm; Goodfellow and McCutcheon 2003).

Ag values for the Nepisiguit deposits range from 10 ppm to 21 ppm and are the lowest Ag values reported from sulphides in the BMC. BMC deposits typically contain Ag values of 50 ppm to 100 ppm, with an average of about 75 ppm (Goodfellow and McCutcheon 2003). Interestingly, the Nepisiguit B sample (Table 1) contains higher concentrations of Ag and Sn than does the Nepisiguit A sample. This difference suggests a larger magmatic component in the former deposit, even though its estimated massive-sulphide tonnage is slightly less than that of the latter deposit.

Rare-earth-element data from massive sulphides can be used to fingerprint the source rocks (mafic or felsic volcanic) and to indicate the physicochemical conditions at the sulphide deposition site. The chondrite-normalized REE profiles for the Nepisiguit A and B massive sulphides (Fig. 18) are typical of massive sulphides found elsewhere in the BMC (Graf 1977; Walker and Lentz, in press). The absolute values of the REEs are lower than those of the host felsic volcanic rocks (Table 1).

The Nepisiguit A and B deposits also have a flat profile or a slightly positive Eu anomaly with respect to the line joining Sm and Gd (Fig. 18). The magnitude of the Eu anomaly may

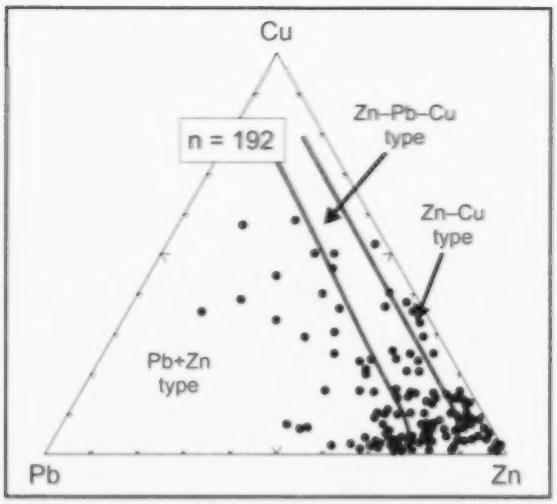

Figure 17. Pb–Cu–Zn ternary diagram for the Nepisiguit A deposit (after Franklin et al. 1981). The assay data are from Stevenson 1955; Stevenson and Brooks 1956; Cominco 1958, 1966; Mitton 1994; and Roy 1996.

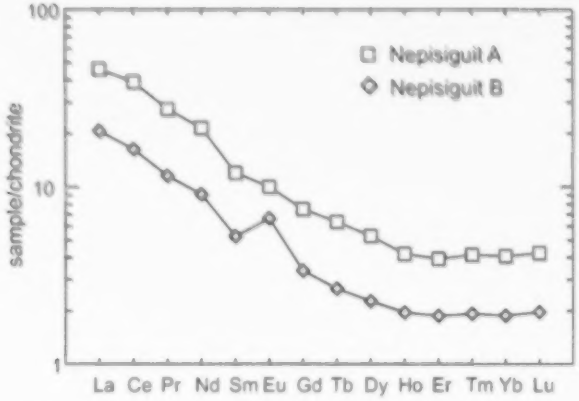

Figure 18. Chondrite-normalized REE profiles for massive-sulphide samples from the Nepisiguit A and B deposits (normalization factors from Nakamura 1974). Note: the Eu anomaly is defined by $Eu_n/Eu^* = Eu_n/\sqrt{(Sm_n \times Gd_n)}$, where *n* represents chondrite normalization.

be helpful in constraining the temperature and pH of the fluids from which the sulphide body formed. This is because Eu^{+3}, like other trivalent REEs, has limited solubility under most hydrothermal and metamorphic pressure and temperature ranges. However, Eu^{+3} is easily reduced to divalent Eu^{+2}, which is much more soluble than Eu^{+3} in mildly to strongly acidic hydrothermal fluids with temperatures higher than 250° C (Sverjensky 1984).

Analyses giving negative Eu anomalies are characteristic of felsic volcanic rocks associated with many of the volcanogenic massive-sulphide deposits in the Bathurst Mining Camp. Eu preferentially fractionates into plagioclase, which settles to the bottom, more mafic, part of the magma chamber (Rollinson 1993). As a result, the fractionated felsic melt becomes depleted in Eu relative to other REEs. If these felsic rocks interact with feldspar-destructive, high-temperature (>250° C), low-pH fluids, minor amounts of REEs will be soluble.

However, the reduction of Eu^{+3} to Eu^{+2} greatly enhances that element's solubility so that its concentration increases in the hydrothermal fluid relative to other REEs. Upon cooling, this fluid will generate exhalitive rocks — that is, massive sulphides — enriched in Eu relative to other REEs (Graf 1977; Gale et al. 1997).

Analysis of the chondrite-normalized REE data from the Nepisiguit A and B deposits (Fig. 18) shows that both samples have negative slopes with elevated LREEs and relatively flat HREEs, a typical profile for massive sulphides derived from fluids reacting with a felsic volcanic footwall or felsic-derived sedimentary sequence. However, the Eu anomalies for these deposits are not large. The Nepisiguit A sample gives an Eu/Eu* value of 1, and the Nepisiguit B sample gives an Eu/Eu* value of 1.7 (Fig. 18). These numbers suggest that temperature and pH of the mineralizing fluids were barely into the field of Eu^{+2} transport, and that the system forming the Nepisiguit B deposit was either slightly hotter, or had a marginally lower pH, than the system forming the Nepisiguit A deposit.

Stable Isotope Studies

Drillhole 94-2 in the western part of the Nepisiguit A deposit passed through the stratiform massive-sulphide lens and into the underlying stringer-sulphide zone; 10 samples from the drill core were collected for $\delta^{34}S$ analysis (Table 2). The five samples from the massive-sulphide lens have an average $\delta^{34}S$ value of +4.94‰, and the two samples from the stringer-sulphide zone have an average $\delta^{34}S$ value of +2.16‰. The three remaining samples, which were collected from the base of the stringer-sulphide zone and the underlying sedimentary rocks, give an average $\delta^{34}S$ value of +4.78‰.

The significantly higher $\delta^{34}S$ value in the massive-sulphide lens relative to that in the stringer-sulphide zone is typical of volcanogenic massive-sulphide systems. This difference can be explained by the greater mixing of contemporaneous ambient seawater with magmatic fluids in the stratiform exhalative horizon, compared with that in the feeder zone. A $\delta^{34}S$ value of about +10‰ to +15‰ has been suggested for Ordovician seawater (as determined from the Selwyn Basin, Yukon), in contrast with typical magmatic $\delta^{34}S$ values of -2‰ to +2‰ (Goodfellow and McCutcheon 2003).

The sulphide samples from drill core 94-2 in the western part of the Nepisiguit A deposit are considerably less enriched in heavier sulphur ($\delta^{34}S$ = +2.1‰ to +4.9‰) than are typical BMS deposits such as the Caribou ($\delta^{34}S$ = +5.9‰ to +9.8‰) in the California Lake Group and Brunswick No. 12 ($\delta^{34}S$ = +12.7‰ to +14.4‰) in the Tetagouche Group (Fig. 19). Massive sulphides from the thickest eastern part of the Nepisiquit A deposit have an average $\delta^{34}S$ of approximately +7.7‰ (Tupper 1960). These values overlap with, but are still lower than, most values reported from deposits in the BMC (Goodfellow and McCutcheon 2003).

Table 2. $\delta^{34}S‰$ values for bulk sulphides from the Nepisiguit A deposit, Bathurst Mining Camp.

Sample Point	$\delta^{34}S‰$	
DDH 94-2 @ 579'	5.04	massive lens
DDH 94-2 @ 580'	6.12	massive lens
DDH 94-2 @ 581.2'	3.19	massive lens
DDH 94-2 @ 583'	4.52	massive lens
DDH 94-2 @ 584'	5.81	massive lens
DDH 94-2 @ 586.5'	1.61	stringer-sulphide zone
DDH 94-2 @ 588'	2.70	stringer-sulphide zone
DDH 94-2 @ 609.5'	4.25	base of stringer-sulphide zone
DDH 94-2 @ 615'	4.67	footwall sedimentary rocks
DDH 94-2 @ 645'	5.42	footwall sedimentary rocks

Analyses carried out by Alison Pye, Memorial University of Newfoundland.

$$\delta^{34}S‰ = \left[\frac{^{34}S/^{32}S\ (\text{sample}) - {}^{34}S/^{32}S\ (\text{standard})}{^{34}S/^{32}S\ (\text{standard})}\right] \times 1000$$

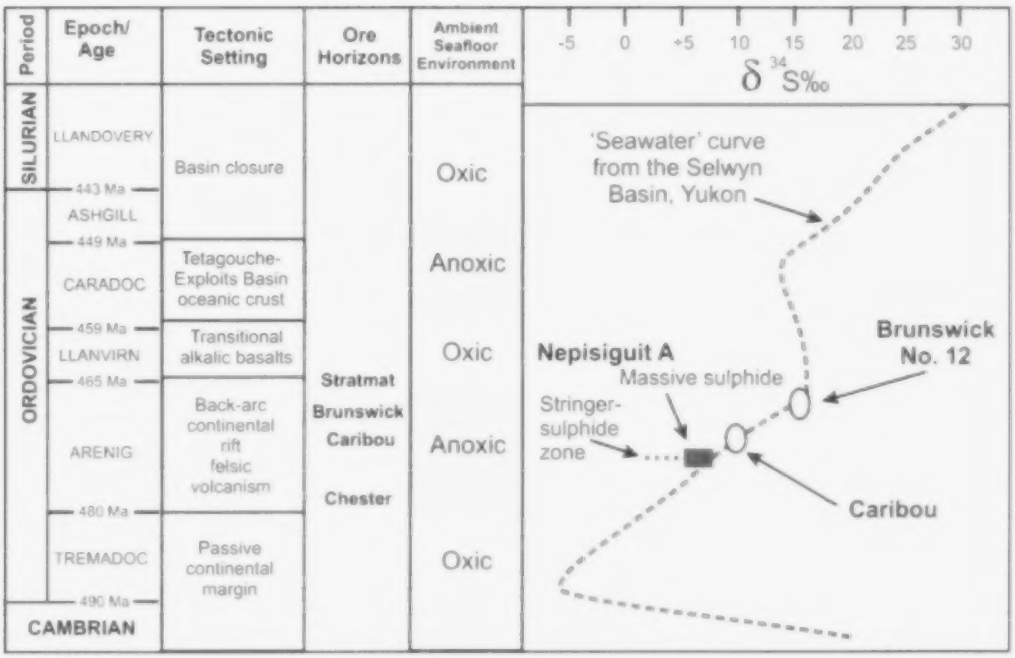

Figure 19. $\delta^{34}S$ data from the Nepisiguit A deposit compared with the 'seawater' curve of the Selwyn Basin, Yukon (modified after Goodfellow et al. 1993 and Goodfellow and McCutcheon 2003).

Most of the sulphur in volcanogenic massive-sulphide deposits of the BMC is of seawater origin (Goodfellow and Peter 1996; Goodfellow 2002) The expectation, therefore, would be that all deposits in the California Lake Group have similar $\delta^{34}S$ signatures because they formed in the same marine subbasin. However, a greater contribution of magmatic sulphur to the venting fluids (that is, lower $\delta^{34}S$ values in the stringer-sulphide zone, slightly higher values in massive sulphides immediately overlying the stringer-sulphide zone, and still higher values in stratiform sulphides distal to the vent site) could explain the anomalously low $\delta^{34}S$ values in the western part of the Nepisiguit A deposit.

Such sulphur isotope variation has been reported at the Brunswick No. 12 deposit (Goodfellow and Peter 1996) and is attributed to metal-laden hydrothermal fluids venting into a stratified water column under reducing (H_2S-dominant) bottom conditions. Thus, the heavier sulphur isotope signature in massive sulphides from the eastern part of the Nepisiguit A deposit can be attributed to their position more distal from the vent site (McCutcheon et al. 1993; Goodfellow and McCutcheon 2003).

Hydrothermal Alteration

The alkali-element values of volcanic rock samples from the Nepisiguit A deposit have already been described under the section on major-element chemistry (see *Lithogeochemistry*). These values lie outside the limits of unaltered volcanic rocks and fall instead within the field of keratophyric rocks (Hughes 1972). Such enrichment in K_2O and depletion in Na_2O for seafloor-altered volcanic rocks has been attributed to alteration at temperatures below 150° C (Galley 1995). Felsic volcanic rocks of the Spruce Lake Formation at the Nepisiguit A deposit have been altered as just described. For this reason, the anomalous alkali values are interpreted to have resulted from normal, low-temperature seafloor alteration rather than from a mineralizing event.

Notwithstanding the above scenario, some minor hydrothermal alteration related to mineralization occurs at the eastern end of the Nepisiguit A deposit, where footwall sedimentary rocks have been slightly bleached (McAllister 1960). Moreover, minor chlorite and significant sericite development in the western part of the deposit are coincident with stringer-sulphide veins that contain pyrrhotite and chalcopyrite as well as pyrite, sphalerite, and minor galena. As can be observed in drill core 94-2, this alteration and mineralization zone immediately underlies the massive-sulphide lens and extends for up to 20 m into the footwall.

Discussion

Inadequate outcrop and lack of drill core both preclude any detailed investigation of stratigraphic positions for the Nepisiguit A, B, and C deposits. The original relationships among these deposits are now obscured because of folding and tectonic interleaving along the thrust fault that juxtaposes the California Lake and Tetagouche groups (Fig. 3). However, the three deposits likely are genetically related, given that they appear within the same small (1.75 km by 0.6 km) area in generally similar host rocks. Clusters of volcanogenic massive-sulphide bodies occurring at one horizon, or multiple lenses occurring at one vent site but at slightly different stratigraphic horizons, both have been recognized in volcanogenic massive-sulphide regions elsewhere such as in the Kuroko district of Japan and the Noranda Mining Camp of Québec (Franklin et al. 1981).

Clusters of volcanogenic massive-sulphide deposits in the Bathurst Mining Camp are situated near specific stratigraphic positions. For instance, the Brunswick No. 12, Brunswick No. 6, and Austin Brook deposits (and several smaller showings) all occur at the same horizon over a strike length of 14 km. As well, multiple sulphide lenses lie within a small area at Taylor Brook and at Heath Steele ACD zones (Fig. 1; McCutcheon 1997; Walker 1999). At the former site, multiple massive-sulphide lenses appear above a single stringer-sulphide zone, whereas at the latter site, sulphide distribution is interpreted to have resulted from the fold repetition of several lenses along a single horizon.

Although the three Nepisiguit deposits are probably genetically related, the precise nature of their relationship is unclear. There are at least three possibilities.

- They may have formed along one exhalite horizon as discrete lenses that were repeated by folding, analogous to what is interpreted for the Heath Steele ACD zones (see Perkins and Roy 1997).
- They may represent the boudinaged remains of a single, attenuated sulphide lens.
- They may have formed as replacement bodies related to one vent site but at slightly different stratigraphic levels, analogous to what is interpreted for the Taylor Brook deposit.

Several questions remain concerning the structure responsible for the distribution of mafic rocks in the Spruce Lake Formation at the Nepisiguit A deposit. Mafic volcanic rocks intersected in drill core at the deposit appear to represent two distinct horizons that occur immediately above and below the sulphide-bearing zone. As described under *Lithogeochemistry*, the footwall mafic sample collected during this study, as well as many of the Noranda samples, are alkalic and unlike the hanging-wall tholeiitic mafic rocks intersected in hole 94-2.

Although the footwall mafic rocks are included here in the Spruce Lake Formation, they more closely resemble other alkali basalts in the Bathurst Mining Camp such as the Brunswick Mines alkali basalt of the Little River Formation in the upper part of the Tetagouche Group; the Camel Back Mountain alkali basalt of the Boucher Brook Formation in the upper part of the California Lake Group; and alkali basalts of the Canoe Landing Lake Formation in the lower part of the California Lake Group (Rogers and van Staal 2003).

Because the footwall sedimentary rocks plot in the Mafic Source field on Figure 13b, they possibly could be assigned either to the younger Little River Formation (Tetagouche Group) or the also-younger Boucher Brook Formation (California Lake Group), instead of to the older, felsic-dominated, Spruce Lake Formation. On the other hand, if these sedimentary rocks are assigned to the Little River Formation, it would be necessary to interpret a cryptic tectonic break (thrust) in the footwall to separate the younger, underlying sedimentary

sequence from the older, overlying felsic volcanic sequence. This interpretation would be problematic, given that all sedimentary rocks in the footwall sequence are somewhat deformed, and that no obvious zone of intensely deformed tectonic mélange was recognized.

On the other hand, if the footwall sedimentary rocks are assigned to the Boucher Brook Formation, then no structural break would be required, because the latter rocks are known to conformably overlie the felsic volcanic sequence of the Spruce Lake Formation (Rogers and van Staal 2003). In this case, however, the rock sequence hosting the Nepisiguit A deposit should young to the south.

The Nepisiguit A deposit was previously thought to young to the south: that is, felsic volcanic rocks lie in the footwall, and sedimentary rocks lie in the hanging wall (Thomas et al. 2000; Goodfellow and McCutcheon 2003). However, the relative positions of the massive sulphides and stringer sulphides within the sericite- to chlorite-altered rocks (interpreted here as representing the feeder zone) suggest that the host sequence in fact is younging to the north. That being the case, the footwall sedimentary rocks and interlayered alkali basalts may be interpreted as laterally equivalent to the Canoe Landing Lake Formation and interfingering with coeval felsic volcanic rocks of the Spruce Lake Formation.

In some respects, the stratigraphic position of the Nepisiguit A deposit is similar to that of the Caribou deposit (Fig. 1), which is hosted by sedimentary rocks overlain by felsic volcanic rocks of the Spruce Lake Formation (Goodfellow 2003). But unlike the Caribou deposit, the Nepisiguit A deposit features mafic rocks capping the sulphides in the immediate hanging wall — a feature that occurs at the Canoe Landing Lake deposit (Fig. 2) situated farther northwest (Walker and McDonald 1995).

As noted previously, no laterally extensive sedimentary exhalite (that is, Algoma-type iron formation) is associated with the Nepisiguit A deposit except for the massive-sulphide deposits discussed in this report. Likewise, Algoma-type iron formation has not been observed elsewhere in the California Lake Group, although metalliferous sedimentary rocks showing evidence of hydrothermal input appear locally along strike from the area of the Nepisiguit deposits (Fig. 14a, 14b). The absence of non-sulphide exhalite at the Nepisiguit A deposit could be explained in one of two scenarios.

- A reducing water column with excess H_2S would preclude iron oxide precipitation and inhibit post-depositional oxidation of any accumulated sulphides.
- A more likely scenario is that eruption of the tholeiitic basalts of the Spruce Lake Formation smothered the hydrothermal system and isolated the sulphide lens from oxidizing effects of the water column.

Given the presence of layered Zn–Pb-rich sulphides adjacent to discordant stringer-sulphide mineralization, it seems reasonable to assign the Nepisiguit A deposit to the proximal autochthonous category of Jambor (1979). Jambor did not classify this deposit as such, because stringer sulphides (the presence of which is critical to the proximal classification) were recognized only later, in deep holes drilled by Noranda (Mitton 1994). It is interpreted that magmatic sulphur (as indicated by low $\delta^{34}S$ values) may have played an important role in forming the stringer sulphides (average $\delta^{34}S$ = 2.1‰) and proximal stratiform massive sulphides (average $\delta^{34}S$ = 4.9‰) in the western part of the Nepisiguit A deposit. On the other hand, the higher sulphur values (average $\delta^{34}S$ = 7.7‰) that characterize stratiform massive sulphides in the more distal, eastern part of the deposit suggest a larger seawater component in that area.

The flat to slightly positive Eu anomalies exhibited by massive sulphides from the Nepisiguit A and B deposits (Table 1, Fig. 18) are typical of volcanogenic massive-sulphide deposits elsewhere and are interpreted to reflect deposition at moderately high temperatures (≥250° C) from moderately low-pH fluids. The interpretation of a moderately high temperature of formation for the Nepisiguit A deposit is supported by its mineral assemblage of chlorite-pyrrhotite-pyrite-chalcopyrite, which occurs in the stringer-sulphide system beneath the massive-sulphide lens.

The presence of sulphide clasts (Fig. 9c) in the hanging-wall fragmental rocks of drillhole 94-2 (Fig. 5a) at 151.2 m (496') is problematic. The clasts are interpreted to represent a primary debris-flow deposit rather than a tectonic mélange. Because these clasts occur at least 130 m above the main sulphide-bearing horizon, they likely are not sourced from the Nepisiguit A deposit. However, at the Wedge deposit situated 9 km west of the Nepisiguit A deposit (Fig. 1), massive-sulphide lenses occur within sedimentary and/or volcaniclastic rocks that directly overlie feldspar-phyric volcanic rocks of the Spruce Lake Formation (Walker and McCutcheon 1996). It is possible that sulphides derived from the Wedge deposit or some other deposit at a similar horizon could be the source of sulphide clasts in hanging-wall rocks of the Nepisiguit A deposit. Therefore, the area underlain by the Spruce Lake Formation north of the Nepisiguit A deposit is a prime exploration target, as the potential exists to discover sulphide deposits at depth in that area, higher in the volcanic-sedimentary section.

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STRATIGRAPHY AND STRUCTURE OF THE MONCTON SUBBASIN IN THE URNEY–WATERFORD AREA, MARITIMES BASIN, NEW BRUNSWICK: IMPLICATIONS FOR THE MCCULLY NATURAL GAS FIELD

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The McCully Natural Gas Field is located in the Moncton Subbasin of the Maritimes Basin near Sussex, southeastern New Brunswick, and was discovered in 2000 by Corridor Resources Inc. and Potash Corporation of Saskatchewan Inc. Seismic surveys and drilling results to date indicate that Mississippian sandstones of the Hiram Brook Member (Albert Formation) hold a gas field of approximately 75 km^2 with an estimated in-place natural gas reserve of 1 trillion ft^3.

In 2005 fieldwork was undertaken in the Urney–Waterford area, approximately 10 km southeast of the McCully Field, to document the sedimentary and tectonic controls on reservoir rocks of the Moncton Subbasin. Carboniferous strata exposed in this area indicate the presence of four unconformities that post-date deposition of the Albert Formation. Seismic and borehole data suggest that structural events related to these unconformities have, to varied degrees, affected the Albert Formation and thus the reservoir sand beds.

Bedrock relationships in the Urney–Waterford area show that the most significant structural event was a Pennsylvanian (post-Mabou Group) northwest–southeast-directed compression. The compression produced thrust movement along the Urney and Clover Hill fault systems, both of which are associated with a hanging-wall anticline and footwall syncline. A subsequent stage of transpression has resulted in a 1 km to 2 km dextral offset together with a strong northeastern deflection of the originally east–west-oriented fold structures. The structural displacement related to the post-Mabou Group transpression phase has largely obscured the older tectonic history in this part of the Moncton Subbasin. However, facies relationships suggest that sedimentation was largely controlled by faulting along basin margins.

Le champ de gaz naturel McCully est situé près de Sussex, dans le Sud-Est du Nouveau-Brunswick, à l'intérieur du sous-bassin de Moncton du bassin des Maritimes. Il a été découvert en 2000 par la Corridor Resources Inc. et la Potash Corporation of Saskatchewan Inc. Les résultats des travaux de forage et des levés sismiques obtenus jusqu'ici révèlent que les grès du Carbonifère inférieur du membre du ruisseau Hiram de la Formation d'Albert abritent un champ gazier d'environ 75 kilomètres carrés représentant des ressources en place estimatives d'un billion de pieds cubes de gaz naturel.

On a réalisé au cours de 2005 des travaux sur le terrain dans le secteur d'Urney-Waterford, à une dizaine de kilomètres au sud-est du champ de gaz McCully, pour documenter les contrôles sédimentaires et tectoniques des roches du réservoir du sous-bassin de Moncton. Des strates du Carbonifère dénudées dans ce secteur révèlent la présence de quatre discordances ultérieures à la sédimentation de la Formation d'Albert. Des données sismiques et des données de forage révèlent que les phénomènes structuraux reliés à ces discordances ont affecté à des degrés divers la Formation d'Albert et en conséquence les lits de sables du réservoir.

Les rapports au sein du substrat rocheux dans le secteur d'Urney-Waterford révèlent que le phénomène structural le plus déterminant a été une compression du Carbonifère tardif (ultérieure à Mabou) s'étant dirigée du nord-ouest au sud-est. La compression a produit un mouvement de poussée le long des systèmes de failles d'Urney et de la colline Clover, tous deux associés à un anticlinal de toit et un synclinal de mur. Un stade subséquent de transpression a entraîné un décalage dextral d'un à deux kilomètres ainsi qu'une déviation prononcée vers le nord-est des structures plissées originalement orientées d'est en ouest. Le déplacement structural apparenté à la phase de transpression ultérieure à Mabou a grandement obscurci le passé tectonique plus ancien dans cette partie du sous-bassin de Moncton, mais les relations des faciès permettent de supposer que la sédimentation a en grande partie été contrôlée par les failles le long des bords du bassin.

INTRODUCTION

The McCully Natural Gas Field (McCully Field) is located approximately 10 km east of Sussex in the southwestern part of the Moncton Subbasin of the Maritimes Basin (Fig. 1, 2). The McCully A-67 discovery well was drilled in 2000 by Corridor Resources Inc. and Potash Corporation of Saskatchewan Inc. (PCS) with two objectives: 1) Corridor's natural gas exploration of the Upper Devonian–Mississippian Horton Group, and 2) PCS's search for a water disposal well for their potash operations (Bertram et al. 2005; Wilson 2005). Boreholes A-67 and P-66 are currently producing approximately 2 million ft^3 of natural gas daily, which is sold to the PCS potash plant for its operations. The McCully Field discovery has opened up a huge potential for exploration plays within the Late Devonian–Carboniferous sedimentary basins of southern New Brunswick and the larger Maritimes Basin (Fig. 2).

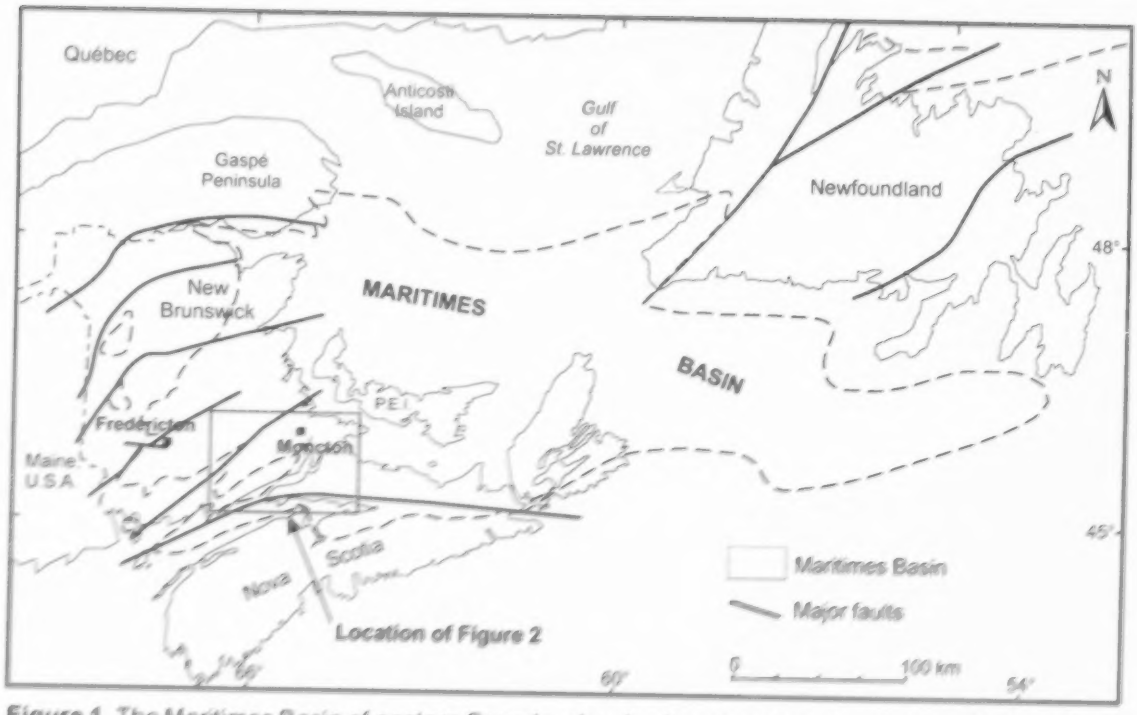


Figure 1. The Maritimes Basin of eastern Canada, showing its present erosional edge. (Modified from Roliff 1962; Williams 1974; Bradley 1982; and Fyffe and Barr 1986.)

The McCully Field reservoir lies at a depth of approximately 2.5 km in tight sandstone intervals of the Mississippian Albert Formation of the Horton Group (Fig. 3, 4). These sand bodies are separated by interbedded siltstone, shale, and organic-bearing shale. The trap is a partly eroded, gentle anticline that is unconformably capped by continental clastic rocks of the Sussex Group. Seismic surveys and drilling results to date indicate that Mississippian sandstones of the Hiram Brook Member of the Albert Formation hold a gas field of approximately 75 km^2 with an estimated in-place natural gas reserve of 1 trillion ft^3.

During the summer of 2005, the New Brunswick Department of Natural Resources (DNR) conducted field investigations south of the McCully Field in the Urney–Waterford area (Fig. 4) to update existing geological maps (see St. Peter 1995a, 1995b, 2002). This report discusses the new surface and subsurface geological data and gives a preliminary tectonic summary of the Urney–Waterford area. The principal goal of hydrocarbon research by the DNR is to document the sedimentary and tectonic controls on reservoir rocks of the Maritimes Basin and to use this information as a basis for assisting present and future exploration efforts.

Regional Geologic Setting

After the Devonian Acadian Orogeny, crustal subsidence of southern New Brunswick and surrounding areas initiated Late Devonian–Carboniferous sedimentation within the Maritimes Basin (Williams and Hatcher 1982; St. Peter 1993). In southern New Brunswick and northern Nova Scotia, the Maritimes Basin consists of the Moncton, Sackville, Cocagne (see St. Peter 2006, this volume), and Cumberland subbasins (Fig. 2). Each subbasin is fully or partly bounded by faults that are interpreted to have been periodically reactivated, possibly

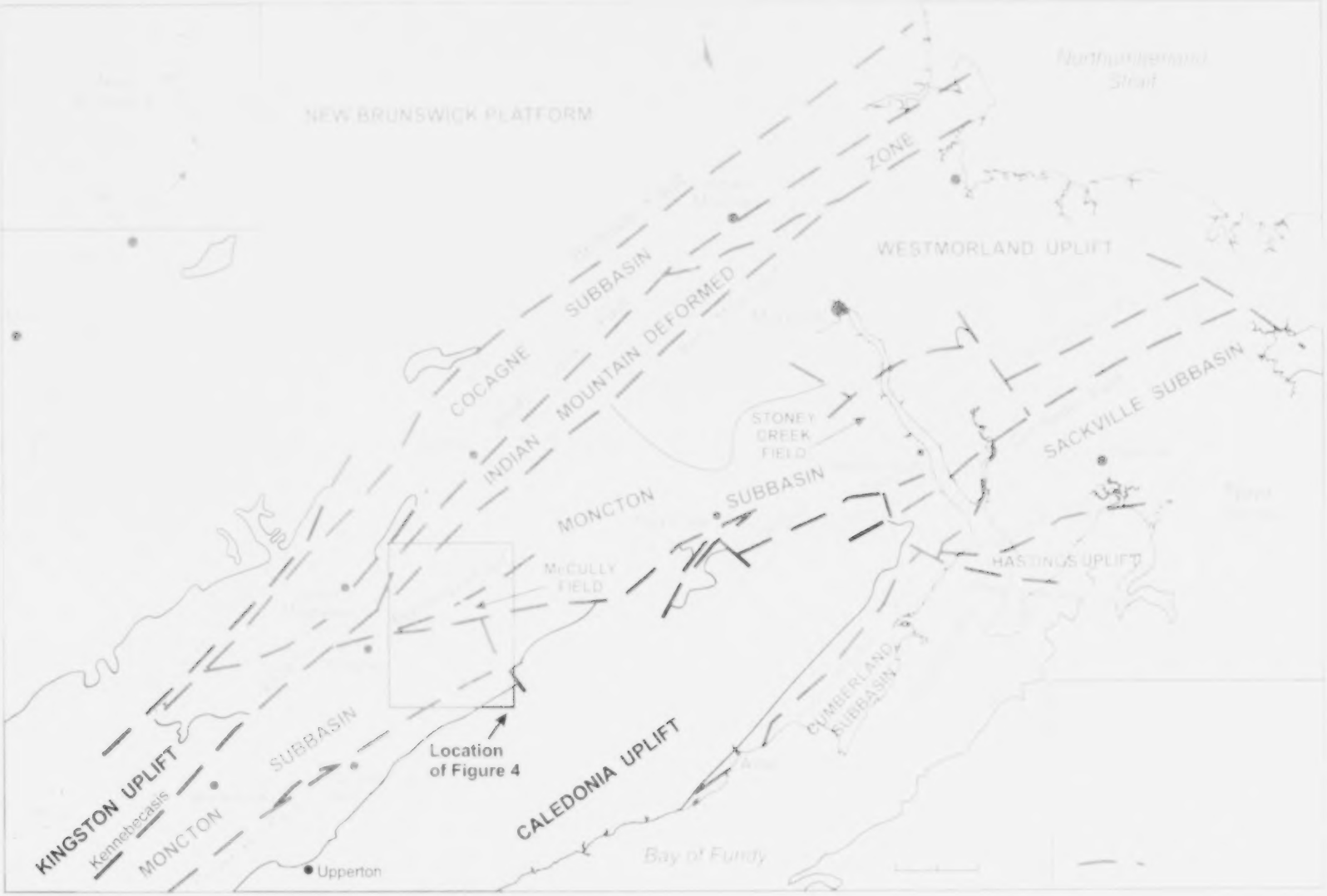

NEW BRUNSWICK PLATFORM
Northumberland Strait
ZONE
WESTMORLAND UPLIFT
COCAGNE SUBBASIN
INDIAN MOUNTAIN DEFORMED
STONEY CREEK FIELD
SACKVILLE SUBBASIN
MONCTON SUBBASIN
McCULLY FIELD
HASTINGS UPLIFT
CUMBERLAND SUBBASIN
Location of Figure 4
KINGSTON UPLIFT
Kennebecasis
MONCTON SUBBASIN
CALEDONIA UPLIFT
Upperton
Bay of Fundy

Figure 2

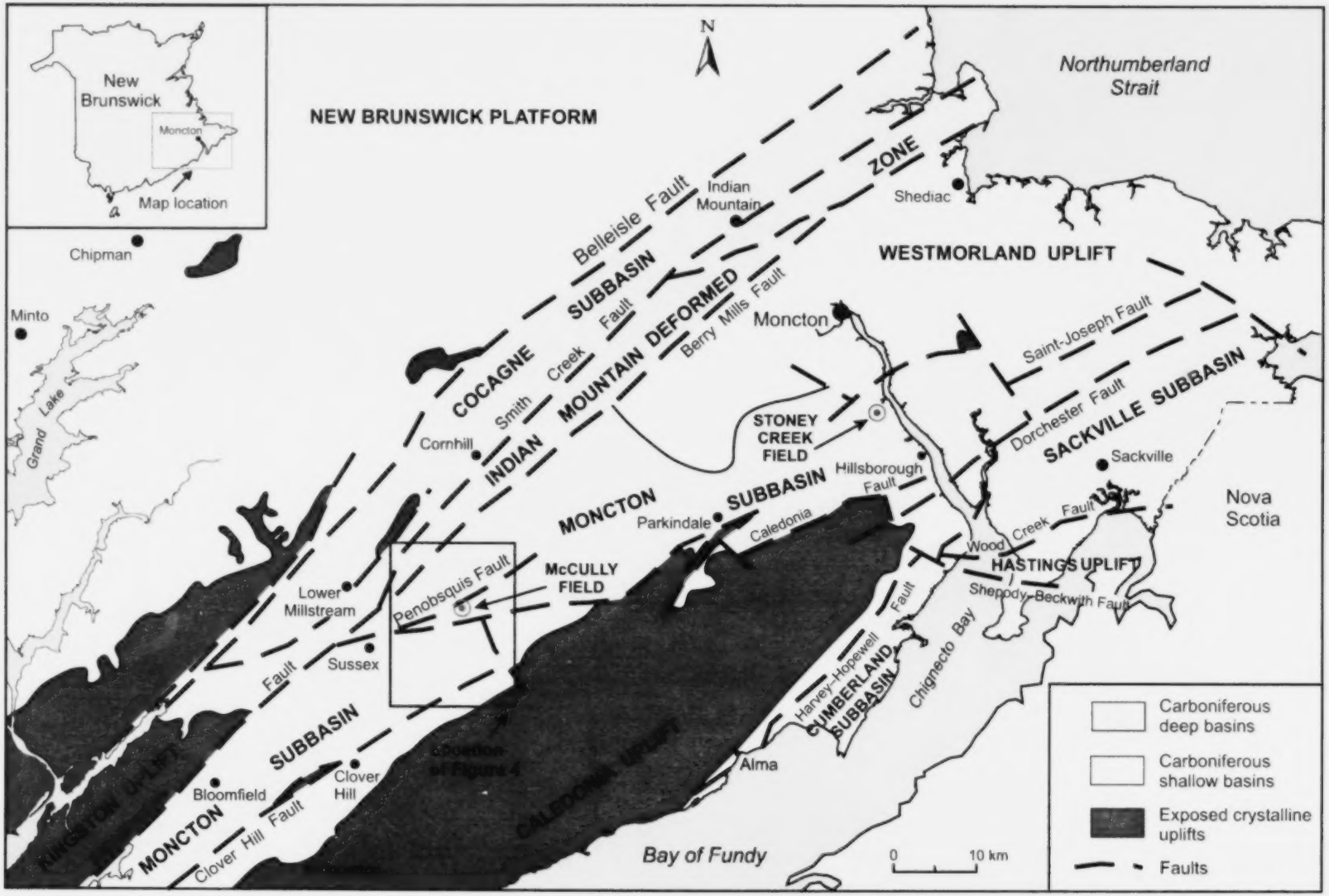


Figure 2. Distribution of subbasins and uplifts in the Maritimes Basin of southeastern New Brunswick (modified after St. Peter 2006, this volume).

several times, during the Carboniferous by a combination of compressional (transpressional) and extensional (transtensional) tectonics (St. Peter 1987; St. Peter 1993; Park and St. Peter 2005). Three exhumed or shallowly buried basement highlands lie immediately adjacent to, or near, the Moncton Subbasin: the Westmorland Uplift to the east, the Caledonia Uplift to the south, and the New Brunswick Platform to the north (Fig. 2)

The crystalline basement that underlies and flanks the Moncton Subbasin in the Urney–Waterford area comprises complexly deformed and metamorphosed sequences of sedimentary, volcanic, and plutonic rocks that range from Neoproterozoic to Lower Silurian (Ruitenberg et al. 1979; Barr and White 1996). The Upper Devonian–Carboniferous section of the Moncton Subbasin that lies on the basement complex is divisible into six groups. From the base up, these are the Horton, Sussex, Windsor, Mabou, Cumberland, and Pictou groups (Fig. 3, 4). Unconformities separate each of these groups, except at the Windsor–Mabou boundary, which apparently is concordant in the central Moncton Subbasin and discordant at the basin edge (St. Peter 2002; Park and St. Peter 2005).

Basement rocks underlying the Caledonian Highlands in the southeast corner of the Urney–Waterford area (Fig. 4, 5) include Neoproterozoic volcanic and sedimentary rocks of the Broad River and Coldbrook groups. The older, Broad River Group comprises mainly intermediate tuffaceous rocks, whereas the younger, Coldbrook Group contains felsic and intermediate tuffs interbedded with volcaniclastic conglomerate, laminated siltstone, and chert (Barr and White 1996, 2004; White et. al. 2001). Layered ultramafic and gabbroic rocks intrude Coldbrook strata northeast of Cedar Camp Brook (Fig. 5). The contact between the Caledonian basement rocks and Upper Devonian–Carboniferous strata is an overstepping angular unconformity (St. Peter 2002; Hinds and St. Peter 2005).

Silurian rocks of the Kingston Group underlie a northeasterly trending ridge northwest of the McCully Field at Jordan Mountain (Fig. 4). There, the Kingston Group comprises interbedded felsic and mafic volcanic rocks and sedimentary rocks intruded by Silurian gabbro, diorite, and granodiorite (Venugopal and Brown 1997). The Silurian rocks at Jordan Mountain are in fault contact with the younger Upper Devonian–Carboniferous strata (Barr et al. 2002; St. Peter 2002; Barr and White 2004).

A system of major northeasterly and minor northerly trending faults transects the Upper Devonian–Carboniferous strata of the Moncton Subbasin (St. Peter 1995a, 1995b, 2002; Hinds and St. Peter 2005; Wilson 2005). The history of movement along these faults shows considerable variation: that is, some faults displace only Horton strata, whereas others displace the entire Carboniferous section. In the Urney–Waterford area, the northeasterly trending faults are associated with hanging-wall anticlines and footwall synclines (Fig. 5, 6). Other phases of tectonism have also affected depositional patterns of Moncton Subbasin strata in the area and are discussed below under *Structural Geology*.

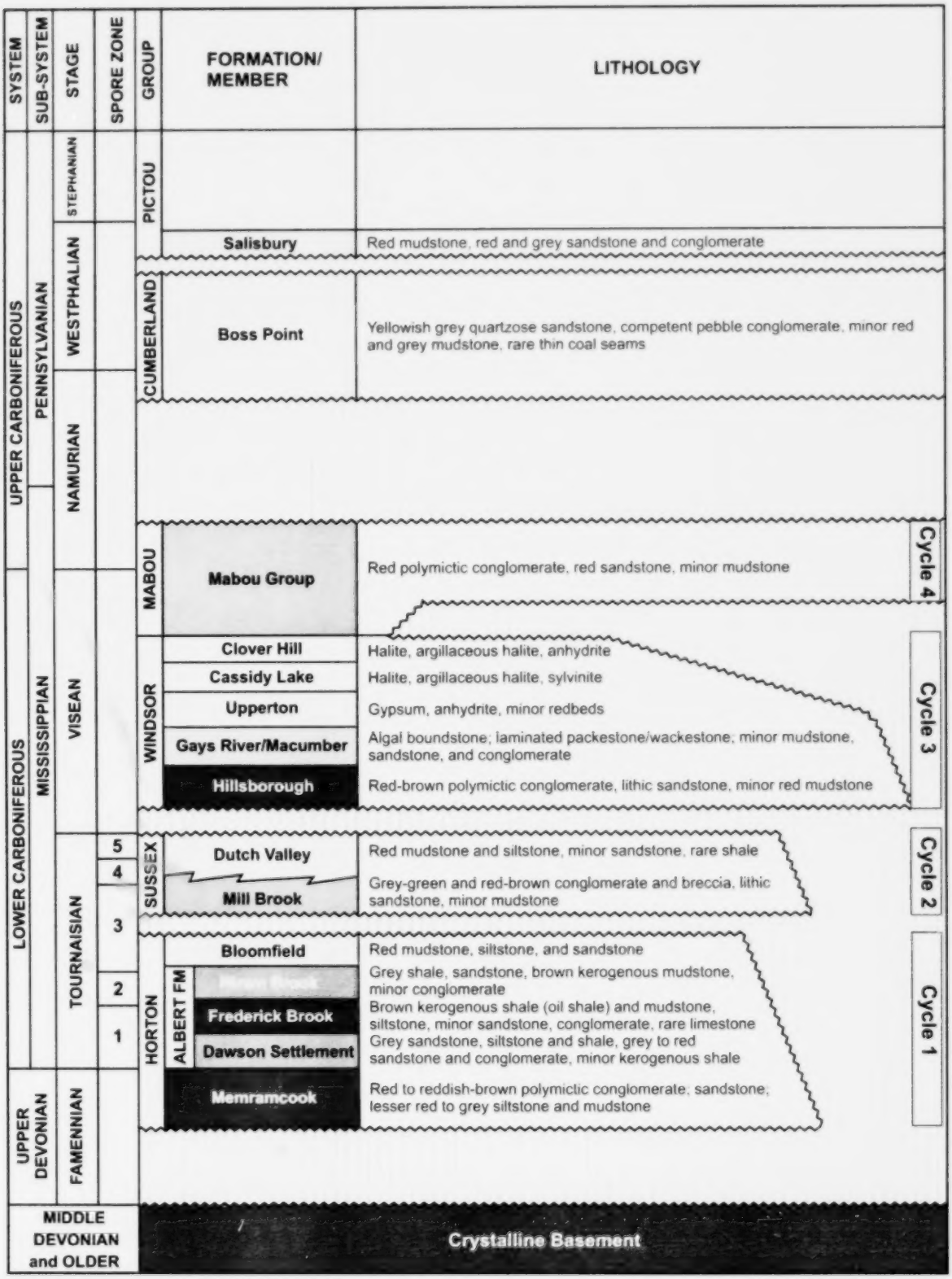


Figure 3. Stratigraphic chart for the Urney–Waterford area, showing stratigraphic units and the four depositional cycles described in the text. (Tournaisian spore zones are after Dolby 2006.)

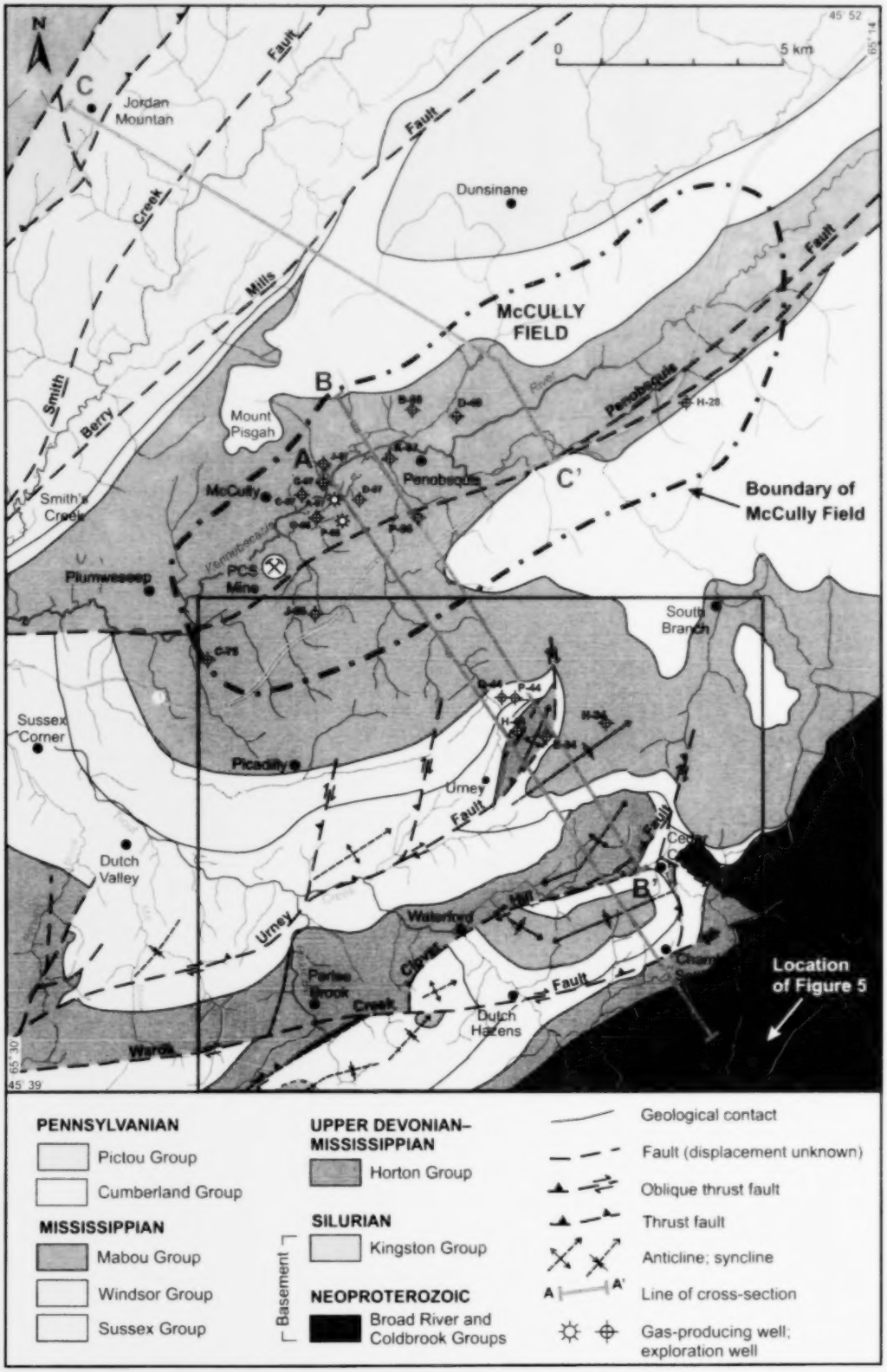
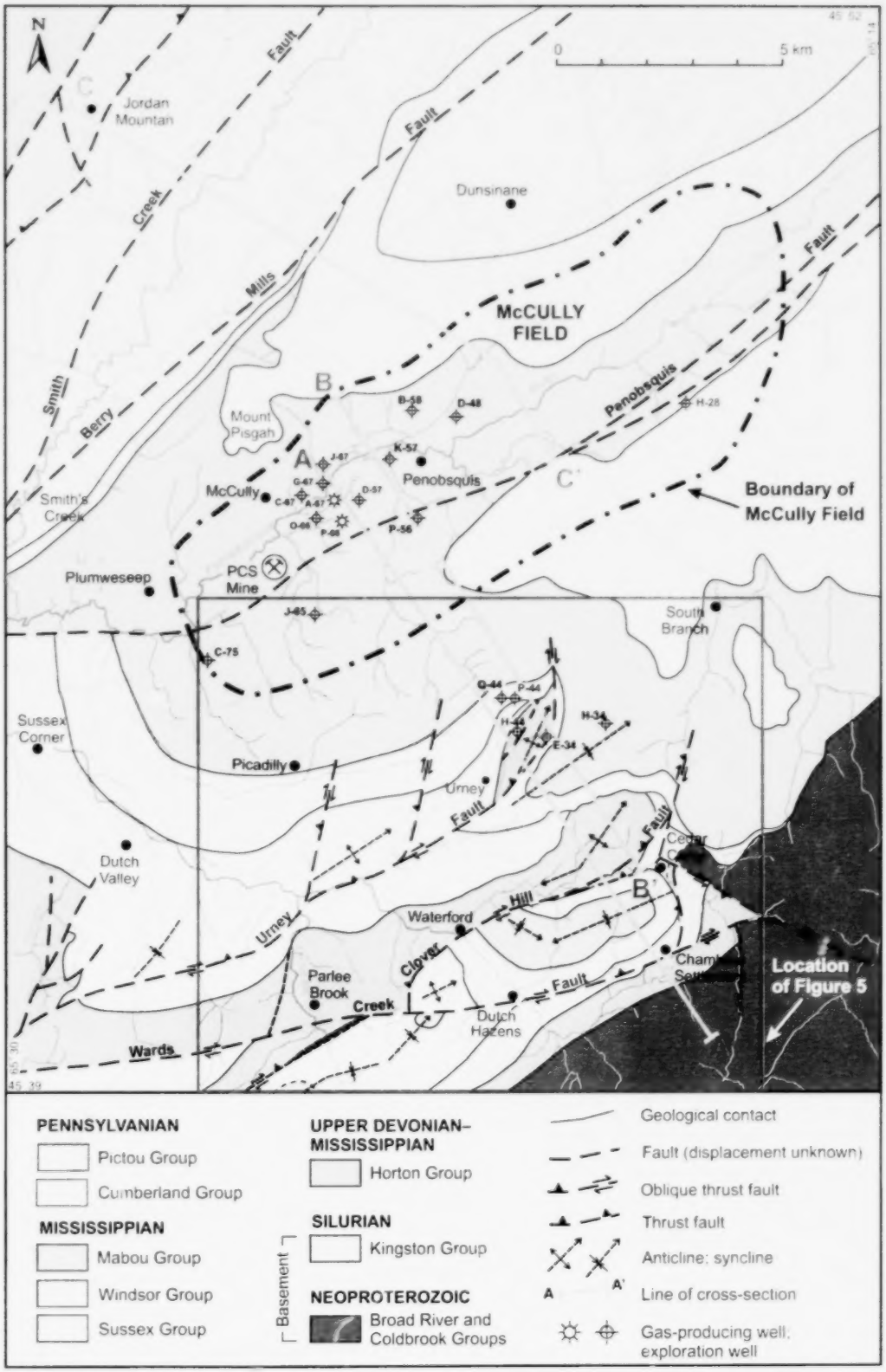


Figure 4. Geology of the McCully Field area. Map shows location of the Urney–Waterford area (Fig. 5) and lines of cross-sections (Fig. 6, 7). See Figure 2 for map location.



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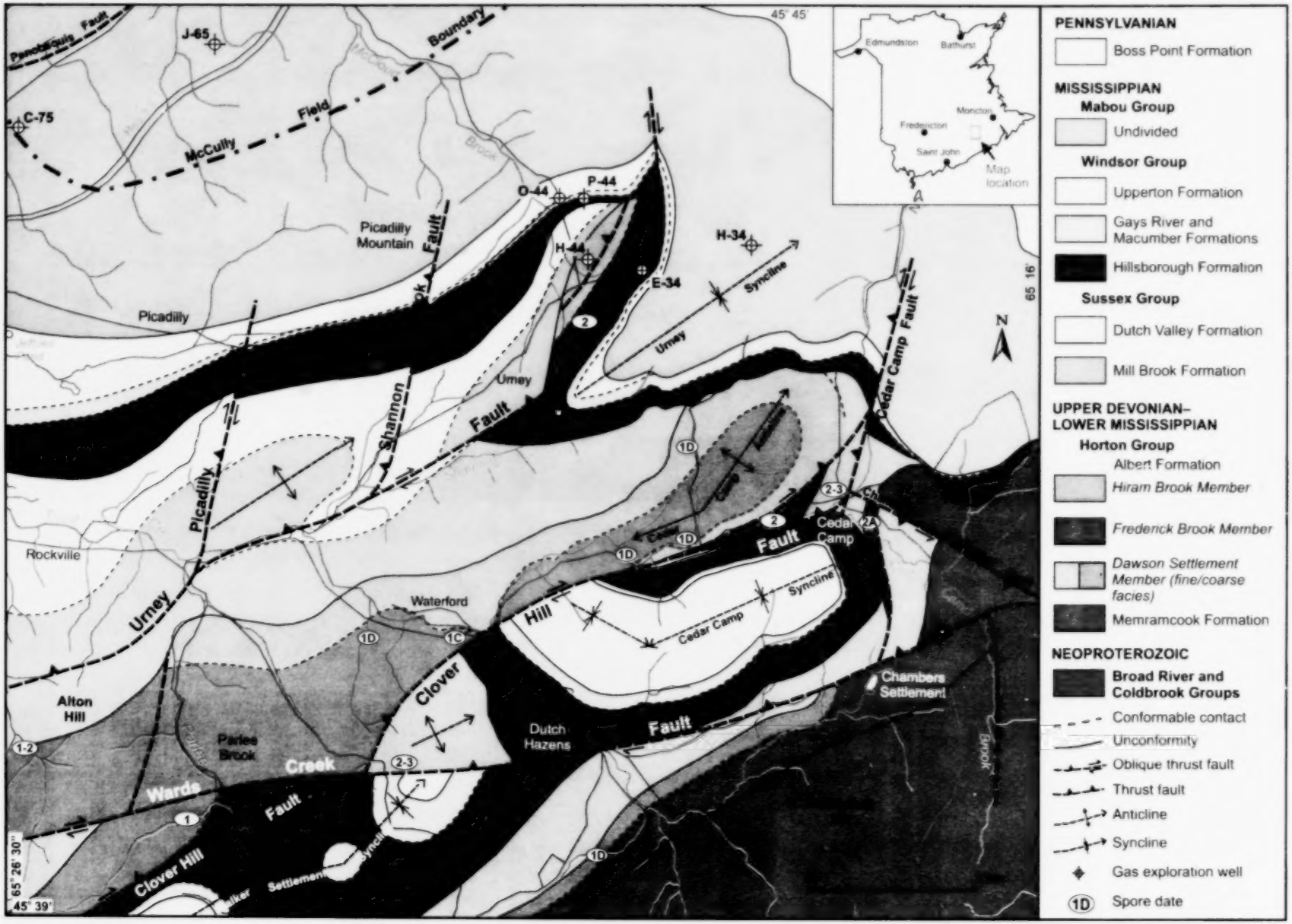

Figure 5. Geology of the Urney–Waterford area, covering parts of NTS 21H/11. See Figure 4 for a more precise map location.

Pennsylvania Knob
J-65
Boundary
C-75
McCully
Field
O-44
H-44
H-34
E-34
Urney Anticline
Sharron Creek Fault
Fault
Urney
Picadilly
Fault
Cedar Camp Fault
Cedar Camp Anticline
Cedar
Chambers Settlement Fault
Hill
Clover
Alton Hill
Paines Brook
McDermott Hill
0
2 km

Figure 5.

UPPER DEVONIAN–CARBONIFEROUS STRATIGRAPHY

Horton Group

In the southwestern Moncton Subbasin, the Horton Group (Fig. 3) comprises a lower red conglomeratic unit (Memramcook Formation), a middle grey shale and sandstone unit (Albert Formation), and an upper red clastic unit (Bloomfield Formation). Its thickness ranges from approximately 250 m in the Urney–Waterford area to more than 1 km in the McCully Field. The Horton strata rest unconformably on the basement complex and are unconformably overlain by the Sussex Group (Gussow 1953; Greiner 1962; St. Peter 1992). The age of the Horton Group in New Brunswick ranges from Late Devonian (Famennian) to Mississippian (Tournaisian) (Greiner 1962; Utting et al. 1989; Dolby 2006).

Memramcook Formation

The Memramcook Formation represents the initial phase of post-Acadian, late Paleozoic sedimentation in the Moncton Subbasin (Carter and Pickerill 1985a; St. Peter 1993). The type section (Norman 1941) is located along Stony Creek, which flows westward into the Memramcook River about 4 km north of the village of Memramcook (Fig. 2). Thickness of the formation ranges from nearly 900 m at the type section (Gussow 1953; Popper 1965) to between 150 m and 250 m in the Chambers Settlement area (Fig. 5).

Purplish red to reddish brown conglomerate and lithic sandstone predominate in the Memramcook Formation and are interbedded with minor red to grey siltstone and mudstone. The conglomerate is generally very poorly sorted with subangular to angular Caledonian Highlands-derived clasts in a poorly sorted, fine- to medium-grained, lithic matrix. Clasts in both the conglomerate and sandstone are commonly coated with desert varnish. The Memramcook Formation is interpreted as being mainly alluvial–fluvial deposits (Greiner 1962; Popper 1965). It ranges from Upper Devonian (Famennian) to Mississippian (early Tournaisian) and is conformable (Fig. 3) with the overlying Albert Formation (Greiner 1962; St. Peter 1992).

Albert Formation

Greiner (1962) subdivided the Albert Formation into three members on the basis of Wright's (1922) stratigraphic model: a lower conglomerate and sandstone unit (Dawson Settlement Member), a middle organic-rich mudstone and shale unit (Frederick Brook Member), and an upper sandstone and shale unit (Hiram Brook Member). Type sections for these three members are located in, and south of, the Stoney Creek Field (Fig. 2). The Albert Formation ranges in thickness from 100 m to 150 m at the subbasin margin to more than 1 km at the McCully Field. Within the map area, the Albert Formation has been significantly eroded by

an unconformity at the base of the Sussex Group. As a result, the upper part of the formation is preserved only at Cedar Camp and northeast of Urney (Fig. 5).

Dawson Settlement Member

The type section for the Dawson Settlement Member is drillhole 166 in the Stoney Creek Field near Dawson Settlement (Greiner 1962). Its distribution and lithology have been determined from surface outcrops in the Urney–Waterford area (St. Peter 1995a, 1995b, 2002) and subsurface seismic profiles of the McCully Field (Bertram et al. 2005). In the Chambers Settlement area along the southern margin of the subbasin, dominant lithotypes are polymictic conglomerate and pebbly sandstone with an estimated total thickness of 150 m to 200 m.

On the northern limb of the Cedar Camp Anticline, the member approaches a thickness of 600 m, comprising a lower coarser facies and an upper finer facies (Fig. 5, 6). The lower facies is dominantly polymictic conglomerate and interbedded sandstone, whereas the upper facies consists of fine to coarse sandstone and lesser siltstone and mudstone. Exposed sections up to 5 m thick near Waterford consist of medium to coarse, poorly sorted sandstone containing about 40% quartz. Bitumen is present in pores of some of these sand beds, and calcite was observed locally as a cementing agent.

In the Urney–Waterford area, the Dawson Settlement Member is interpreted to have been deposited in distal braided-stream, fan-delta, and lacustrine-margin environments. About 2 km south of Sussex, its upper contact is gradational with the fine-grained Frederick Brook Member (St. Peter 2002). Spore assemblages for the Dawson Settlement Member generally indicate an age of Tournaisian Zone 1A to 1B, but they also suggest an age as young as Zone 1D in the Urney–Waterford area (Fig. 3, 5; Dolby 2002, 2006).

Frederick Brook Member

The type section for the Frederick Brook Member is located along Frederick Brook, about 10 km south of the Stoney Creek Field (Fig. 2). The member is at least 800 m thick, as determined from seismic and borehole intersections (Fig. 6). Within the Urney–Waterford area, Frederick Brook strata are exposed along the southern limb of the Cedar Camp Anticline and along the northwestern limb of the Urney Anticline (Fig. 5, 6). Dark brown to greyish brown, kerogen-rich, dolomitic, laminated mudstone and shale are characteristic of the unit. Siltstone, sandstone, and limestone beds are present as minor components (Greiner 1962; St. Peter 1992). In the Urney and Cedar Camp areas, the upper part of the Frederick Brook Member has been partly eroded beneath the unconformity with the Sussex Group (St. Peter 1995a).

The Frederick Brook Member has the highest overall organic matter content of the three members of the Albert Formation and is considered the primary source rock for gas of the

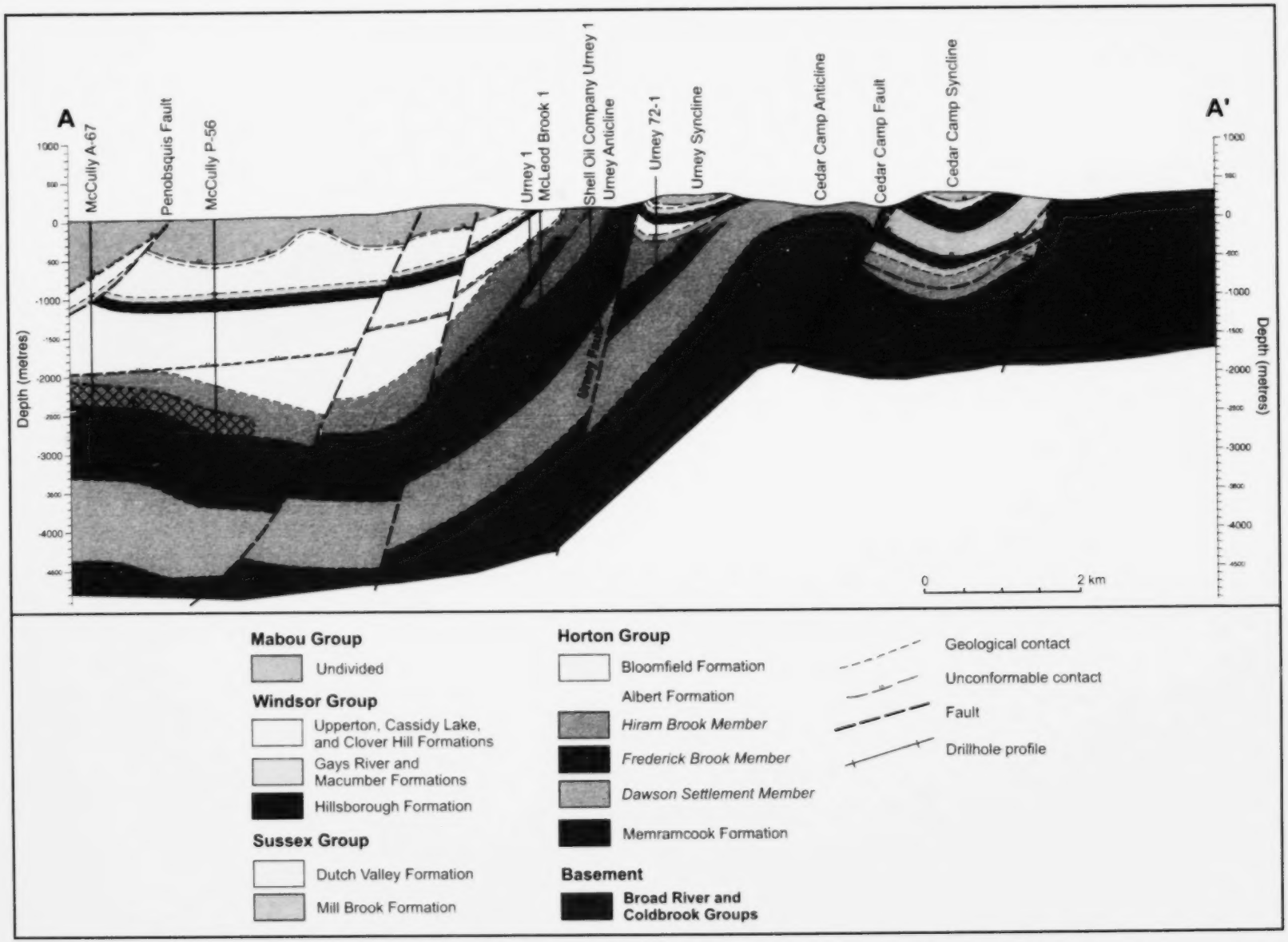

Figure 6. Regional cross-section A–A' through the Urney–Waterford to McCully areas. See Figure 4 for line of cross-section.

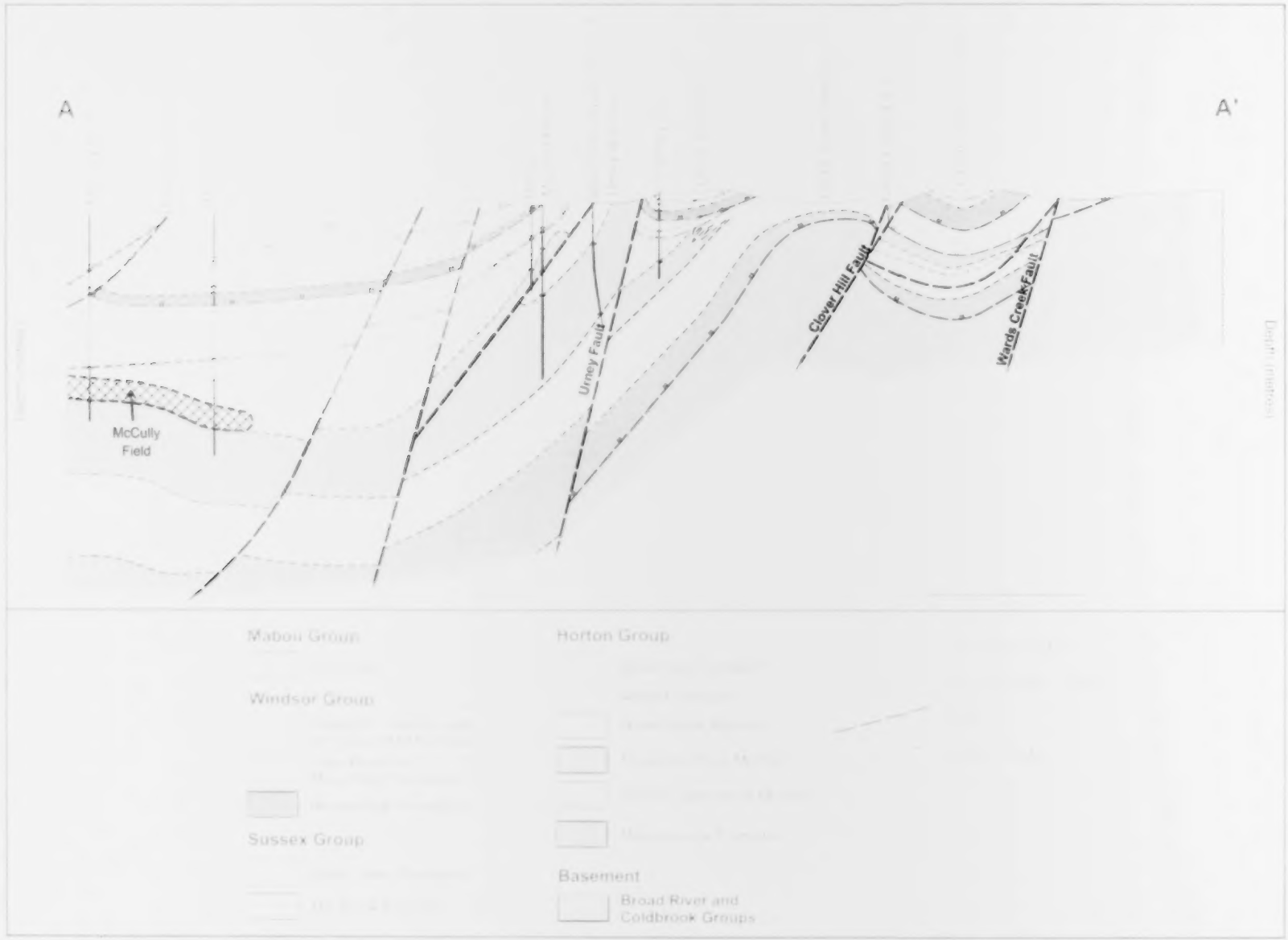

Figure 6 Regional cross-section A–A' through the Urney, Waterbury to McCully areas. See Figure 4 for line of cross-section.

McCully Field (Bertram et al. 2005). Toward the top of the section, the organic shale becomes progressively siltier and grades into the overlying siltstone and sandstone of the Hiram Brook Member (Wright 1922; Greiner 1962; St. Peter 1989). Albertite (a black, solid, vitreous bitumen) occurs locally as veins within the kerogen-rich, dolomitic, laminate facies of the Frederick Brook Member. In most places the veins are only a few millimetres wide, but at Albert Mines near Frederick Brook, a vein up to 18 m was mined from the Frederick Brook Member during the mid-19th century (St. Peter and Johnson 1997; Martin 2003). The albertite veins have formed mostly in situ or in very close proximity to their source (Farrell and Mossman 1995).

The depositional setting of the Frederick Brook Member is interpreted as a shallow to deep lacustrine sequence (Greiner 1962; Carter and Pickerill 1985b; St. Peter 1992). Spore assemblages within and adjacent to the McCully Field area generally indicate an age of Tournaisian Zone 1C to 1D for the Frederick Brook Member (Fig. 3; Dolby 2002), but they may indicate an age as young as Zone 2 in the Urney–Waterford area (Fig. 5; Dolby 2006).

Hiram Brook Member

The type section of the Hiram Brook Member is defined in the New Brunswick Oilfields well 104 in the southwestern part of the Stoney Creek Field (Fig. 2; Greiner 1962). In the Urney–Waterford area, Hiram Brook strata are exposed on the northwestern limb of the Urney Anticline (Fig. 5). They also occur in the subsurface northeast of there, hosting gas-bearing sands of the McCully Field (Fig. 6).

The Hiram Brook Member comprises an interbedded sequence of grey shale, siltstone, and sandstone with lesser beds of kerogenous mudstone and conglomerate. An exposure on the west side of the Urney Anticline comprises light brown, parallel-bedded sandstone interbedded with swaley-trough, cross-stratified and ripple-laminated sandstone. The swaley intervals locally carry pebbles, are approximately 50 cm thick, and may represent channel lenses. On the northwestern limb of the Urney Anticline and in the subsurface immediately southeast of the Urney Fault, the member is about 500 m to 600 m thick. The Hiram Brook section has been partially eroded beneath the unconformity at the base of the Sussex Group in the Urney Anticline. Farther south, the section has been completely cut out by the same unconformity (Fig. 5). In the subsurface northeast of Urney, the Hiram Brook Member is succeeded by the Bloomfield Formation (Fig. 6).

The presence of swaley cross-bedding and wave-ripple-laminated sandstone interstratified with kerogenous, laminated mudrock suggests that the Hiram Brook Member was in large part deposited in a shallow lacustrine environment. However, fining-upward sandstone sequences identified in well records at the McCully Field show that at least some of the sand bodies represent fluvial deposits. Spore assemblages indicate an age of Tournaisian Zone 2A to possibly as young as early Zone 3A for the Hiram Brook Member (Fig. 3; Dolby 2002).

Bloomfield Formation

The type section for the recently designated Bloomfield Formation (St. Peter 2006, this volume) is situated near Bloomfield, about 25 km southwest of Sussex (Fig. 2). The section consists of a redbed sequence that gradationally overlies interstratified kerogenous mudstone and sandstone of the Albert Formation. The formation comprises grey to green and reddish brown shale and mudrock in its lower part; pink, coarse feldspathic sandstone and polymictic conglomerate in its middle part; and reddish brown mudrock and minor sandstone in its upper part.

In the Urney–Waterford area, the Bloomfield Formation appears only in the subsurface as a preserved wedge beneath the basal unconformity of the Sussex Group. A 400 m thick interval of these strata was intersected in borehole J-65 (Fig. 4, 5) below the basal conglomerate marker of the Mill Brook Formation (Sussex Group) (Fig. 6). The Bloomsfield section in the borehole consists of interbedded reddish brown and grey sandstone, siltstone, and shale that grade downward into the predominantly grey sandstone, siltstone, and shale of the Hiram Brook Member.

As noted above, the Bloomfield Formation concordantly overlies the Albert Formation in its type section. However, some seismic profiles suggest the Bloomfield redbeds may lie disconformably upon the Albert Formation in the McCully Field (Corridor Resources Inc., unpublished data). Those same profiles leave little doubt that the upper boundary of the Bloomfield wedge with the Sussex Group (Fig. 6) is an angular unconformity (also see Wilson 2005). The Bloomfield Formation has not been dated from the McCully Field wells, but spore assemblages from the Bloomfield and Lower Millstream areas (Fig. 2) indicate a Tournaisian Zone 2 or 3A age (Fig. 3; St. Peter 2006, this volume).

Sussex Group

Drilling of the first McCully Field exploration borehole A-67 and subsequent analysis of new seismic data by staff of Corridor Resources Inc. led to the discovery that an unconformity existed within strata of the Horton Group, as then defined. The thick sequence of redbeds immediately above the unconformity was named the Sussex Group by St. Peter (2002). These beds were previously included in the Weldon Formation in the uppermost Horton Group (McCutcheon 1978; St Peter 1989).

In the Urney–Waterford area, the Sussex Group comprises coarse clastic rocks of the Mill Brook Formation and fine clastic rocks of the Dutch Valley Formation (St. Peter 2002). The Sussex Group is unconformably overlain by the Windsor Group (St. Peter 2002; Hinds and St. Peter 2005).

Mill Brook Formation

The Mill Brook Formation comprises the coarse redbed strata that lie unconformably above the Albert Formation from Sussex Corner east to Urney (Fig. 4, 5). The type section is located along Mill Brook (Fig. 4) from the Albert Formation in the south to the fine redbeds of the Dutch Valley Formation in the north. The Mill Brook Formation is lithologically equivalent to the coarse basal facies of the Weldon Formation in the Hillsborough area (Norman 1941; McCutcheon 1978).

The Mill Brook Formation is composed of greyish green to reddish brown, polymictic, pebble to cobble conglomerate with minor sandstone and mudstone interbeds, the latter increasing in frequency toward the top of the section. The formation is about 400 m thick near Dutch Hazens in the Urney–Waterford area (Fig. 5), but 12 km farther north, in the subsurface of the McCully Field near the basin centre (Fig. 6), it thins to between 10 m and 50 m. On the northern flank of the Cedar Camp Anticline, the formation is gradationally succeeded by the fine facies that are characteristic of the Dutch Valley Formation. On the southern flank, the Mill Brook beds are unconformably overlain by conglomerate of the Hillsborough Formation of the Windsor Group (St. Peter 2002; Hinds and St. Peter 2005).

The Mill Brook Formation is interpreted as an alluvial-fan complex, due to the presence of 1) massive, poorly sorted, thick beds of polymictic conglomerate believed to be mass flows, and 2) associated intervals of more thin-bedded, rounded, polymictic clast conglomerate and sandstone that likely represent high-energy braided stream deposits (Bull 1972; Leeder 1982). Both the wedge shape of the formation (thick in the south and thinning toward the north) and the similarity of its clasts to Caledonia Uplift lithotypes strongly suggest a southern provenance. Attempts to recover spores from the Mill Brook Formation have been unsuccessful, but spore assemblages in coarse basal beds of the Sussex Group in the Indian Mountain area (Fig. 2) yield Tournaisian Zone 3B and younger ages (St. Peter 2006, this volume).

Dutch Valley Formation

The fine redbeds referred to as the Dutch Valley Formation (St. Peter 2002) are equivalent to those placed in the upper part of the Weldon Formation by McCutcheon (1978). Scattered outcrops along the stream bed of Trout Brook and its tributaries in the Dutch Valley area have been designated as the type section (Fig. 4). The unit, which extends from Dutch Valley eastward to Rockville and from there to McLeod Brook in the Urney–Waterford area (Fig. 5), consists mainly of red siltstone, mudstone, and shale with minor interbeds of red lithic sandstone and rare beds of polymictic conglomerate. The fine redbeds contain locally abundant calcrete nodules that in places coalesce to form nodular limestone beds.

As previously noted, the Dutch Valley Formation gradationally overlies the Mill Brook Formation. Its upper contact in the Urney–Waterford area is a disconformity to angular unconformity at the base of the Hillsborough Formation (Fig. 5, 6). For example, beds of the Dutch Valley Formation are overstepped in the Urney Anticline and are absent beneath the Windsor Group in the Cedar Camp Syncline.

The thickness of the Dutch Valley Formation shows significant variation. In the type area it is about 450 m, but beneath the unconformity at the base of the Windsor Group northeast of Urney it thins out entirely (Fig. 5). North of the type area, the thickness increases to at least 900 m over the McCully Field (Fig. 6). The presence of pedogenic calcrete and locally observed mudcracks in the fine-grained red lithotypes suggests an arid or semi-arid, distal alluvial-fan or broad fluvial-floodplain environment. Given this interpretation, the northward thickening of the Dutch Valley Formation and the concurrent thinning of the Mill Brook Formation seem to indicate that the formations represent a proximal–distal alluvial-fan pair.

No palynomorphs have been recovered from the Dutch Valley Formation in the present map area. However, in the eastern and northeastern Moncton Subbasin, spore assemblages in stratigraphically equivalent beds of the Gautreau and Weldon formations yield Tournaisian Zone 3B to 5 ages (St. Peter 2006, this volume).

Windsor Group

Deposition of the carbonate formations of the Windsor Group represents the initiation of open marine conditions that extended over an area of approximately 90,000 km^2 from Newfoundland through Nova Scotia to southern New Brunswick (McCutcheon 1981). Within the western portion of the Moncton Subbasin, the Hillsborough Formation of the Windsor Group represents a basal, coarse, continental clastic facies conformably overlain by the shallow to deeper marine, mainly limestone, facies of the Gays River and Macumber formations. Upper formations of the Windsor Group are composed mainly of sulphate and chloride facies (Gussow 1953; McCutcheon 1981; St. Peter 1992).

Contact relationships between the various formations of the Windsor Group are interpreted to range from conformable to interfingering (McCutcheon 1981; St. Peter 1992). Outcrop evidence throughout the Moncton Subbasin indicates that the Hillsborough Formation was deposited unconformably on top of the basement and the Sussex and Albert groups (Gussow 1953; Hinds and St. Peter 2005). The unique appearance of the limestone sequences of the Macumber and Gays River formations serves as a marker horizon across most of the Moncton Subbasin.

Hillsborough Formation

The coarse redbeds in the type area of the Hillsborough Formation near Hillsborough, which lies about 20 km south of Moncton (Fig. 2), have been described by Wright (1922), Norman (1941), and Gussow (1953). In the Urney–Waterford area, the Hillsborough Formation forms prominent ridges and low hills from the north side of Trout Creek in the west to Urney in the east, and from Parlee Brook in the southwest to Cedar Camp in the northeast (Fig. 5). From where it strikes east onto the crystalline basement, the formation is folded around the Urney Anticline and Urney Syncline. Regionally, the Hillsborough Formation forms a wedge-shaped body that ranges in thickness from approximately 300 m at Dutch Hazens to 200 m at Picadilly, thinning out entirely over the McCully Field (Fig. 5, 6).

This formation consists of reddish brown to locally grey or olive green, polymictic, pebble to boulder conglomerate, conglomerate sandstone, and minor sandstone. The clasts vary from angular to subrounded and comprise a wide range of volcanic, plutonic, and sedimentary rock types. The thick, massive, poorly sorted beds are almost certainly debris flows (Bull 1972; Leeder 1982), and the thick-bedded, clast-supported conglomerates with a sand matrix are likely braided-stream deposits. The Hillsborough Formation appears to represent an alluvial-fan complex with a southern provenance, as evidenced by the formation's wedge shape, the presence of mass-flow and waterlain deposits, and the abundant clasts resembling lithotypes of the Caledonia Uplift.

Regional field relationships have convincingly demonstrated that the boundary between the Hillsborough Formation and the Sussex Group varies from a disconformity to an angular unconformity. The formation lies in more-or-less horizontal disposition with the underlying Dutch Valley beds at Trout Creek in the west, but progressively cuts down-section to eventually sit on crystalline basement east of Cedar Camp (Fig. 5). The Hillsborough Formation is structurally concordant with the overlying marine carbonates of the Gays River and Macumber formations (McCutcheon 1981). The unconformity between the Hillsborough Formation and underlying Sussex Group, and the concordant boundary of the formation with the overlying marine strata, together provide sound justification for assigning the Hillsborough Formation to the depositional cycle of the Windsor Group (Wright 1922).

No dateable palymorph assemblages have been collected from Hillsborough beds. Given its stratigraphic position and the observed hiatus above the latest Tournaisian Dutch Valley Formation of the Sussex Group (Fig. 6), the Hillsborough Formation is likely early Visean.

Gays River and Macumber Formations

Distribution of the early Visean Windsor carbonates in the Urney–Waterford area (Fig. 5) has been outlined by McCutcheon (1981). Limestone of the Gays River Formation is

dominantly a thick to massive, bedded, algal boundstone with minor, thin-bedded wackestone and packstone. The Macumber Formation, in contrast, consists of parallel-laminated to thin-bedded wackestone to packstone. The Macumber limestone varies from grey to black, is moderately organic-bearing, and typically emits a petroliferous odour when freshly broken. Fine-laminated beds of the Macumber Formation indicate deposition in deeper portions of the subbasin, whereas algal mounds and reef facies of the Gays River Formation are interpreted to be lateral equivalents deposited in shallower water along the subbasin margins (McCutcheon 1981; Schenk et al. 1994).

The Gays River Formation overlies the Hillsborough Formation on the southern flank of the Cedar Camp Anticline. The Gays River Formation is thickest (about 35 m) in the Cedar Camp Syncline and thins northwards toward the Urney Anticline to an average of 5 m. A poorly defined facies change of the Gays River Formation to the Macumber lithofacies occurs in the north around the Urney Anticline and west to Sussex Corner (Fig. 5). Thickness of the Macumber Formation ranges from 1 m to 5 m on the southeastern limb of the Urney Syncline and increases to 20 m to 30 m in wells in the McCully Field (Fig. 6). Limestones of the Gays River and Macumber formations are conformably overlain by evaporites of the Upperton Formation.

Upperton Formation

The type section of the Upperton Formation is located near Upperton, 28 km southwest of Sussex (Anderle et al. 1979; McCutcheon 1981). The formation outcrops rarely in the Urney–Waterford area, and its surface distribution is largely inferred from karst topography. An analysis of cross-section A–A' (Fig. 6) suggests a formational thickness of 40 m to 80 m.

The Upperton Formation is composed mainly of replacive and displacive anhydrite and gypsum facies that occur in an original host of greenish grey mudstone and laminated limestone. The formation has a conformable relationship with the overlying Cassidy Lake Formation (Anderle et al. 1979; McCutcheon 1981). Spores from stratigraphically equivalent beds from a borehole near Stewiacke, Nova Scotia, indicate a middle Visean age for the Upperton Formation (Utting 1980).

Cassidy Lake Formation

The type section of the Cassidy Lake Formation occurs in borehole intersections from the Clover Hill potash deposit (Anderle et al. 1979; McCutcheon 1981), situated 25 km southwest of the McCully Field (Fig. 2). The formation is composed mainly of halite and sylvinite evaporite facies, the latter of which forms the ore zone of the potash deposit being mined at Plumweseep by Potash Corporation of Saskatchewan (PCS) Inc. (Fig. 4).

In the subsurface at the PCS mine and above ground at the McCully Field, the total thickness of halite and sylvinite facies ranges from nonexistent to approximately 400 m (Fig. 6, 7). The variation is due to both salt doming processes and structural deformation (Roulston et al. 1995; Wilson 2005). The distribution and stratigraphic thickness of the Cassidy Lake evaporites suggest deposition in the central part of a deep, density-stratified marine basin. Algal buildups in the Gays River Formation likely were responsible for the restriction of marine circulation that led to evaporite participation (McCutcheon 1981).

Borehole observations suggest that the contact between the Cassidy Lake Formation and overlying Clover Hill Formation is conformable with some indications of dissolution (Anderle et al. 1979; McCutcheon 1981; Wilson 2005). Spore assemblages reported by McCutcheon (1981) from the Cassidy Lake evaporites yield an Assemblage Zone 1 (Utting 1980) or middle Visean age (Fig. 3).

Clover Hill Formation

Like the Cassidy Lake Formation, the Clover Hill Formation is present only in the subsurface; the type section occurs in borehole intersections at the Clover Hill potash deposit. The Clover Hill Formation consists mainly of anhydrite and halite facies, with rare interbedded shale, and its thickness in the subsurface ranges from 10 m to 20 m (Anderle et. al. 1979; McCutcheon 1981; Wilson 2005). The Clover Hill evaporites are clearly marine deposits, but the gradational contact with terrestrial redbeds of the overlying Mabou Group indicates an upward shallowing from deep-water evaporites of the underlying Cassidy Lake Formation (Anderle et al. 1979; McCutcheon 1981).

Seismic profiles suggest that the Clover Hill Formation and overlying Mabou Group are conformable near the basin axis of the McCully Field (Fig. 6). Furthermore, the anhydrite facies of the Clover Hill Formation intersected in Corridor Resources and PCS boreholes near the basin centre contain red-brown mudstone interbeds resembling those in the overlying Mabou Group, indicating that the contact is gradational (Wilson 2005). The contact becomes an unconformity at the basin edge, where the overlying Mabou Formation cuts the underlying stratigraphy and rests along the basement on the eastern edge of the Urney–Waterford area (Fig. 5).

The Clover Hill Formation is interpreted to have been deposited mainly in an arid, nearshore–intratidal environment (McCutcheon 1981). Stratigraphic relationships suggest that the age of the Clover Hill Formation is middle Visean (McCutcheon 1981). Spore assemblages from the uppermost Windsor beds elsewhere in New Brunswick, although not specifically assigned to the Clover Hill Formation, have also been assigned a middle Visean Assemblage Zone 1 age (Utting 1980).

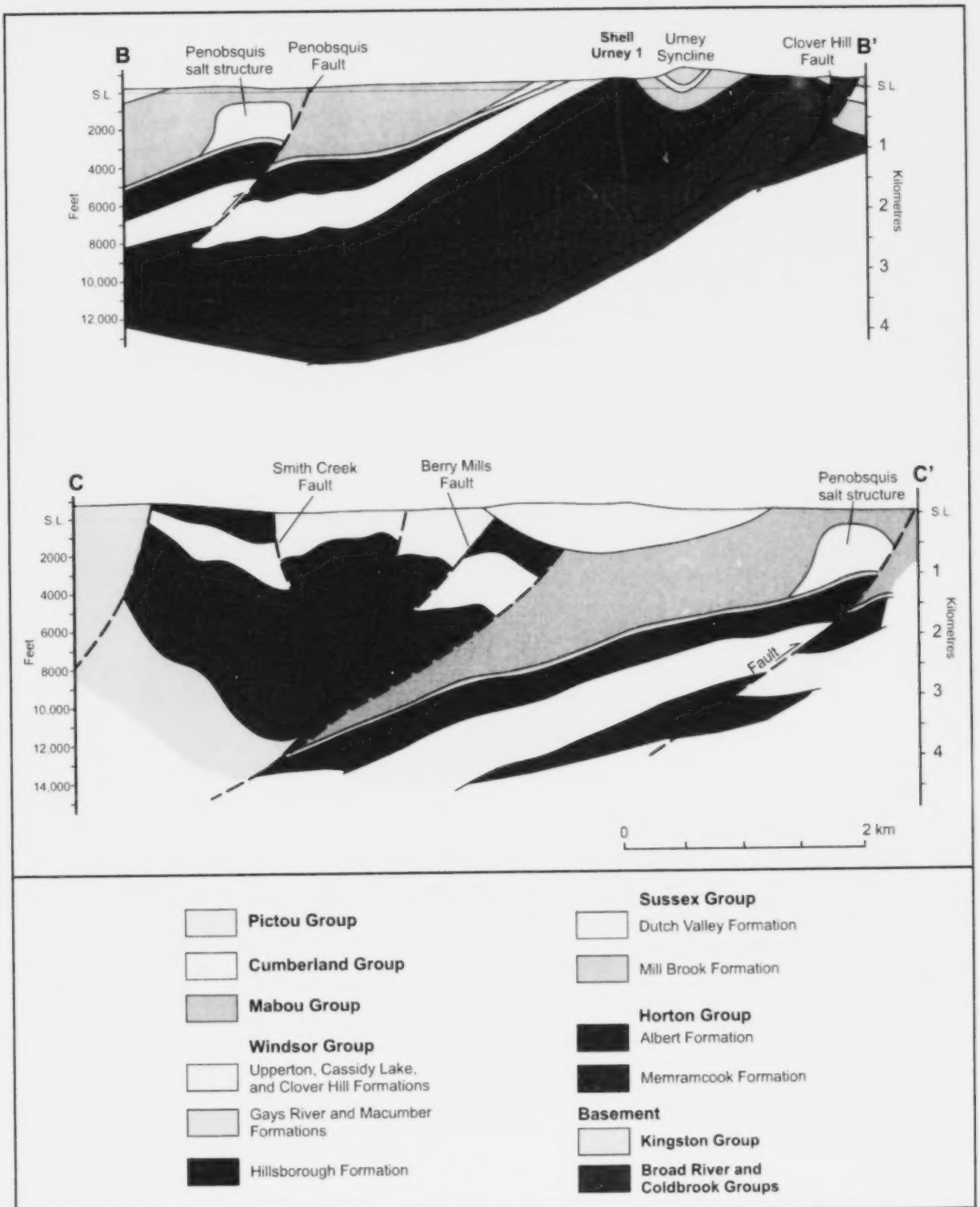

Figure 7. Regional cross-sections B–B' and C–C' in the McCully Field area (after Gussow 1953). See Figure 4 for lines of cross-sections.

Mabou Group

The Mabou Group (formerly the Hopewell Group) in New Brunswick comprises red, polymictic conglomerate, sandstone, and mudstone. This redbed sequence varies greatly in thickness from more than 1 km in the Cumberland Subbasin (Ryan and Boehner 1994) to less than 50 m on the flanks of the Caledonia Uplift, Westmorland Uplift, and New Brunswick Platform (St. Peter, unpublished data). The sequence is approximately 600 m thick in the central portion of the southwestern part of the Moncton Subbasin, as determined in the subsurface of the McCully Field at borehole P-56 (Fig. 4, 6).

The Mabou Group in the Urney–Waterford area has not been divided into formations during this present investigation. On the western end of the Cedar Camp Syncline, outcrops exhibit 30 cm to 50 cm thick, planar- and cross-stratified conglomerate and medium- to coarse-grained, pebbly sandstone interbeds, which indicate deposition in a high-energy, probably braided-stream, environment. Spore assemblages from beds in the eastern Moncton Subbasin, Sackville Subbasin, and western Cumberland Subbasin indicate a late Visean to early Namurian age for the Mabou Group (St. Peter and Johnson 2005). In the Mount Pisgah area (Fig. 4), redbeds of the Mabou Group are overstepped unconformably by grey strata of the Cumberland Group (St. Peter 2002).

Cumberland Group

In the Urney–Waterford area, the Cumberland Group is represented by yellow-grey, quartz-rich sandstone and interbedded quartz-pebble conglomerate of the Boss Point Formation (Gussow 1953; St. Peter 2001). Surface mapping of the McCully Field area by St. Peter (2002) and seismic interpretations by Wilson (2005) indicate the Boss Point Formation reaches a maximum thickness of about 300 m and extends continuously across central portions of the Moncton Subbasin. The contact of the Boss Point Formation with the underlying redbeds of the Mabou Group in the Moncton Subbasin is disconformable to unconformable with sharp lithological and paleontological boundaries (St. Peter and Johnson 1997). Spore assemblages indicate that the Boss Point Formation ranges in age from late Namurian to early Westphalian B (St. Peter 2001; St. Peter and Johnson 2005).

Pictou Group

In the McCully Field area, the Pictou Group (Fig. 4) is represented by the Salisbury Formation. The type section of the formation is situated along the Petitcodiac River and its tributaries in Westmorland County, New Brunswick (St. Peter 2001). The formation is composed mainly of reddish brown mudstone and siltstone with lesser interbeds of sandstone and pebbly sandstone, and minor thin coal seams (Stewart 1941). The sandstone-siltstone-mudstone beds commonly occur in stacked, fining-upward cycles. The

sandstone varies from lithic to arenitic and often is characterized by trough cross-bedding. Northeast of Dunsinane, the thickness of the formation exceeds 100 m (St. Peter 2001; Wilson 2005). The unconformity at the base of the Pictou Group extends down-section as far as the Sussex Group in the McCully area and to basement in the northeastern Moncton Subbasin (St. Peter and Johnson 1997; St. Peter 2001). Spore assemblages from the Salisbury Formation range in age from late Westphalian B to Westphalian D (St. Peter and Johnson 2005).

STRUCTURAL GEOLOGY

Previous Work

Descriptions of structural geology in the Moncton Subbasin can be traced back to the early 1800s (Gussow 1953). The first major mapping survey in the Urney–Waterford area was conducted by the Geological Survey of Canada between 1940 and 1945 (Evans and Alcock 1945; Alcock and MacKenzie 1946). This survey ascribed the tectonic mechanism causing the northeasterly trending fold structures in the Urney–Waterford area to compression after an initial period of basin subsidence and sediment deposition. One thrust fault was recognized in the area (Evans and Alcock 1945) and was later named the Clover Hill Fault (Fig. 6) by Gussow (1953).

Gussow (1953) used seismic sections to interpret the Clover Hill and Berry Mills faults within the Urney–Waterford area as being related to a compressional mode of deformation (Fig. 5). His cross-sections illustrate mostly single-thrust faults and related folding near the Caledonian Highlands and more imbricate thrust systems, which resemble partial flower structures, near the basin centre just northeast of the McCully Field (Fig. 7). He was the first worker to attribute major unconformities such as those at the base of the Windsor and Cumberland groups to separate tectonic events within the Moncton Subbasin.

Since the 1960s, the generally accepted model of fault displacements in the Moncton Subbasin has been transpression associated with the oblique continental collision of the North American and African plates during the late Palaeozoic (Williams and Hatcher 1982). In a transpressional regime, several episodes of extensional and compressional tectonics can occur in the same area during the process of oblique collision (Hobbs et al. 1976).

In southeastern New Brunswick, the variation in tectonic modes throughout the Late Devonian and Carboniferous caused alternating events of basin subsidence and uplift. Each stratigraphic group in the study area represents a depositional cycle (Cycles 1 to 4 on Fig. 3) that consisted of subsidence followed by uplift, erosion, and a resultant unconformity. In the Urney–Waterford area, the intersecting fault patterns can be related, in part, to tectonic

dynamics that gave rise to the unconformities (St. Peter 1995a, 1995b, 2002) that are described as follows (see Fig. 3):

- a mid-Tournaisian unconformity separating the Horton and Sussex groups (marking the end of Cycle 1),
- an earliest Visean unconformity separating the Sussex and Windsor groups (marking the end of Cycle 2),
- a late Visean unconformity to disconformity separating the Windsor and Mabou groups (marking the end of Cycle 3), and
- a mid-Namurian unconformity separating the Mabou and Cumberland groups (marking the end of Cycle 4).

Cycle 1

During the Late Devonian, oblique extension is interpreted to have generated mainly normal faults within basement rocks of the Maritimes Basin. The down-dropped blocks then infilled with sediments of the Memramcook Formation and equivalent formations within the Maritimes Basin. In southeastern New Brunswick, the major subsiding blocks formed the Moncton, Cocagne, and Sackville subbasins bounded by basement highlands of the New Brunswick Platform, and the Westmorland, Hastings, and Caledonia uplifts (Fig. 2). The topographic relief must have been steep at times during deposition of the Memramcook and Albert formations, as evidenced by debris flows, soft sediment deformation, and slump structures observed in these strata throughout the Moncton Subbasin (St. Peter 1992, 1993; Wilson 2005). In southern New Brunswick, the subsiding basins were terrestrial and apparently had no opening to the sea (Greiner 1974).

The existence of relic normal faults in the Urney–Waterford area can be interpreted in cross-section A–A' (Fig. 4, 6) that was constructed from boreholes and industry seismic data (Bertram et al. 2005). All faults identified in this section dip steeply north-northwest. The Clover Hill Fault farther south has also been identified in the subsurface (Fig. 7) from older seismic surveys (Gussow 1953). This fault demonstrably exerted a significant influence on the depositional and tectonic environment/history along the southern margin of the Moncton Subbasin throughout the Late Devonian to Pennsylvanian.

Surface extrapolation of the stratigraphy supports the interpretation that a downward slip resulted in the deposition of a 1 km thick sequence of Horton strata within blocks enclosed by the Wards Creek and Urney faults (Fig. 6). Borehole intersections show that the thickness of the Horton Group increases in a northerly direction to approximately 3 km along a system of three steeply dipping faults, including the Urney Fault. This suggests that structures such as the Urney Fault originally were part of a system of down-to-the-west

normal faults during the late Famennian to early Tournasian, despite the reverse-fault or thrust-fault displacement now apparent at the Urney Anticline (Fig. 5, 6; St. Peter 2002).

In the middle Tournaisian, the tectonic environment changed from extensional to dominantly transpressive, resulting in the cessation of normal-fault movement and the initiation of compressional folding of both lithified and semi-lithified rocks of the Horton Group (Park and St. Peter 2005). As well, greater thicknesses of the Horton strata were removed near the basin margins through uplift and erosion of the Fredericks Brook and Hiram Brook members. As a result, the Dawson Settlement Member is unconformably overlain directly by rocks of the Sussex Group in the Cedar Camp Syncline.

Surface mapping and the seismic section through the Urney–Waterford area indicate that the Horton strata were folded to some extent before deposition of the more gently dipping Sussex Group. On the northwestern flanks of the Urney and Cedar Camp anticlines, bedding-dip angles between the Horton and the overlying Sussex strata vary by as much as 25°. Compressive forces related to this post-Horton–pre-Sussex Group deformation evidently were more north–south-directed, in contrast with the southeast–northwest-oriented forces required to generate later northeasterly trending structures such as the Urney Syncline. These early fold structures likely extended laterally for at least 10 km, roughly parallel to the margin of the Caledonia Uplift. All normal faults in the area could have been reactivated during the Tournaisian compressional event, but the steep dip angle of most faults would have limited the reverse motion (Fig. 6).

In the eastern Moncton Subbasin, exposures of the fine-grained Albert facies display complex internal folding and faulting structures that pre-date deposition of the Windsor Group (Park and St. Peter 2005). This process occurs in stratigraphy composed of incompetent shale sequences with only thin (1–50 cm) siltstone to sandstone interbeds sandwiched between thick, competent sandstone and conglomerate packages. In general, the shale facies of the Frederick Brook Member in the core of the Urney Anticline display similar high-amplitude fold structures, while the competent conglomerate facies of the Dawson Settlement Member display low-amplitude folds west of Cedar Camp. The small-scale fold structures can be inferred from borehole data and are impossible to discern in seismic sections. Nonetheless, they collectively could accommodate large horizontal displacement across the basin such as has been postulated for the eastern Moncton Subbasin (Park and St. Peter 2005). Some of the thicker fine-grained sequences of the Albert Formation at McCully Field could be tectonically influenced in this fashion.

Palynological evidence indicates that the brief period of oblique compression and associated uplift during Cycle 1 was restricted to the middle Tournaisian (Fig. 3). The middle Tournaisian unconformity marking the end of this cycle is readily observed in the Urney–

Waterford area, from the central McCully Field to the basin margin (Hinds and St. Peter 2005; Wilson 2005).

Cycle 2

The second depositional cycle (Fig. 3) began with renewed subsidence along most or all of the relict normal faults in the southwestern part of the Moncton Subbasin, leading to deposition of the Sussex Group. Redbeds of the Sussex Group were laid down on folded Horton strata during a period of oblique extension during the middle to late Tournaisian. Seismic profiles indicate that reflectors in the Sussex Group are mainly undisturbed, suggesting that little internal deformation took place during the uplift that occurred before deposition of the Windsor Group in the earliest Visean. However, some low-amplitude folding of the Sussex Group and older strata is evidenced by the 10° to 15° dip difference in outcrop between the Mill Brook Formation and overlying Hillsborough Formation along the northern flank of the Cedar Camp Anticline.

Cycle 3

The early Visean marks a major phase of extension that resulted in deposition of continental clastic rocks of the Hillsborough Formation and younger marine carbonates of the Windsor Group. In the middle Visean, subsidence appears to have waned as evaporite facies of the Cassidy Lake and Clover Hill formations were deposited. In the late Visean, a renewed period of compression resulted in uplift of the basin margins and partial erosion of the Windsor section (Fig. 3, 5). In the deeper, more central parts of the basin, the effects of uplift were less apparent: in some locations, such as near the basin axis of the McCully Field, the Mabou Group lies conformably on the uppermost member of the Windsor Group. Thus, the boundary between the Windsor and Mabou groups appears to be conformable in deeper parts of the basin, indicating that the degree of uplift decreased away from the basin margins.

Cycle 4

In the late Visean, extensional tectonics initiated deposition of redbed sequences of the Mabou Group. Two salt domes appear as structural highs in seismic profiles north of Urney, the northern one being host to the PCS potash deposit (Fig. 4, 7). These highs indicate that evaporite facies of the Windsor Group had begun to deform and flow in the late Visean to early Namurian as a result of burial beneath the accumulating Mabou sediments. Subsurface isopach maps portray the Penobsquis 'salt high' (Fig. 7) as a continuous ridge about 2 km wide and extending at least 15 km in a northwesterly direction (Wilson 2005). Wilson et al. (2006) present a detailed description of salt tectonics and deformation in the Windsor Group.

This cycle of uplift and partial erosion affecting the Mabou Group and older Upper Devonian–Carboniferous units represents the most widespread compressional tectonic event in the geological evolution of the Maritimes Basin. The middle Namurian unconformity related to this event occurs across the entire Maritimes Basin and marks the cessation of locally derived redbed accumulation that dominated deposition in the Horton, Sussex, and Mabou groups, which together define the Moncton Subbasin. Following the regional peneplanation, widespread deposition began of the distally derived, quartz-rich sandstone and conglomerate of the Cumberland Group (Boss Point Formation).

Deformation related to the middle Namurian uplift is evident along some faults in the Urney–Waterford area. Significant oblique movement occurred along these faults, as indicated by the exposure of basement slivers along the Clover Hill Fault in the Shannon Brook and Cedar Camp areas (Fig. 5, 6). A greater component of thrust movement is apparent at the eastern ends of these faults, where compressive stresses reorientated the Clover Hill and Wards Creek faults from a northeasterly into a more northerly direction against a basement high (Fig. 6). Both faults are associated with hanging-wall anticlines (Urney and Cedar Camp anticlines) and footwall synclines (Urney and Cedar Camp synclines) that demonstrate the northerly dip of these major faults. The footwall deformation on the Urney Fault is clearly observed east of Urney. Here, steeply dipping limestone of the Macumber Formation displays microfolds and brecciated horizons, whereas conglomerate of the Mabou Formation situated farther from the fault trace dips less steeply and is less penetratively fractured (Fig. 4, 6). Bedding dips of the Horton and Windsor groups reach 85° within a 200 m area on both sides of the eastern end of the Urney Fault.

Stereonet analysis also indicates that oblique movement along the Clover Hill Fault refolded the axial trace of the Cedar Camp Anticline and Cedar Camp Syncline (Fig. 6). This bending was the result of stress accommodation along a splay at the eastern end of the Clover Hill Fault. The compressional bending also created a wedge structure between the Cedar Camp Fault and the splayed Clover Hill Fault (Fig. 6). This wedge is marked by steeply dipping beds in close proximity to the main Clover Hill Fault and its subsidiary splay, and by calcite veining in conglomerate of the Mill Brook Formation just west of Cedar Camp. Strata of the Horton and Sussex groups in this wedge have been thrust over carbonates of the Windsor Formation in the Cedar Camp Syncline.

Stress accommodation likely was transferred to the Wards Creek Fault farther west, as the deflection angle of the Clover Hill and Cedar Camp faults reached a critical angle. The renewed movement of the Wards Creek Fault offset the Clover Hill Fault and the Cedar Camp–Walker Settlement synclinal structure. As a result of this late fault movement, the Cedar Camp Syncline axis was refolded into a doubly plunging structure.

The easterly displacement of the Cedar Camp fault block was further accommodated by easterly movement along an unnamed fault on the eastern tip of the syncline. This westerly dipping thrust fault positions older strata of the Horton Group over younger rocks of the Sussex Group. The fault zone is marked by a highly strained zone of calcite-filled fractures in redbeds of the Mill Brook Formation along Cedar Camp Brook. A second high-strain zone about 1 km farther north on Cedar Camp Brook has a more easterly trend and marks the intersection of the unnamed thrust and Chambers Settlement Fault (Fig. 5). Conglomerates of the Memramcook and Hillsborough formations east of Cedar Camp exhibit steeply dipping beds near these faults. Additional displacement of the Cedar Camp fault block likely was accommodated by movement along the Chambers Settlement Fault at the contact between the resistant basement block and less resistant Carboniferous basin rocks.

Two relatively late fault structures west of Urney, those being the Shannon Brook and Piccadilly faults, formed as minor splays off the main Urney Fault (Fig. 5). Conglomerate of the Hillsborough Formation has a more northerly trend in the immediate area of the Shannon Brook Fault but displays an easterly trend away from the fault. These faults likely would have accommodated stresses built up from the northward deflection of the eastern termination of the Urney Fault.

CONCLUSIONS

- The succession in the southwestern part of the Moncton Subbasin is divided into four depositional cycles, represented by the Horton, Sussex, Windsor, and Mabou groups. The cycles are separated by major unconformities along basin margins or by correlative conformities in deeper parts of the basin.
- Unconformities that define group boundaries are the consequence of associated folding and block-faulting. Periodic reactivation of these faults, particularly the Clover Hill Fault, was responsible for the initiation and cessation of sedimentation in each cycle.
- Strata previously included in the upper part of the Horton Group are assigned to the newly designated Sussex Group. The newly defined Horton and Sussex groups are separated by a middle Tournaisian angular unconformity.
- The Sussex Group comprises an alluvial–fluvial fan complex consisting of a southern, coarse, proximal wedge (Mill Brook Formation) and a northern, fine, distal wedge (Dutch Valley Formation).
- Surface oil or bitumen shows are present in rocks of the Horton, Sussex, and Windsor groups in the Waterford, Cedar Camp, and Urney areas.
- A commercial gas resource is present and under development in the McCully Field. The reservoir is in lacustrine–fluvial sandstones of the Hiram Brook Member of the Albert Formation. The trap is an anticline truncated and sealed by several hundred metres of mudstones of the unconformably overlying Dutch Valley Formation (Sussex Group).

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GEOLOGICAL RELATIONSHIP BETWEEN THE COCAGNE SUBBASIN AND INDIAN MOUNTAIN DEFORMED ZONE, MARITIMES BASIN, NEW BRUNSWICK

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The Cocagne Subbasin of southeastern New Brunswick is a Late Devonian–Early Carboniferous mainly terrestrial basin within the composite late Paleozoic Maritimes Basin of eastern Canada. A stratigraphic and structural study of the Cocagne Subbasin was undertaken to determine its relationship with bounding crystalline basement rocks of the New Brunswick Platform to the northwest and the Indian Mountain Deformed Zone (IMDZ) to the southeast. The IMDZ is a faulted complex of Proterozoic basement rocks and Carboniferous strata and is flanked on its southeastern side by the Moncton Subbasin and Westmorland Uplift.

The Late Devonian–Mississippian basin fill in the Cocagne Subbasin is analogous to that in the Moncton Subbasin and comprises four groups: the Upper Devonian–Mississippian Horton Group, and the Mississippian Sussex, Windsor, and Mabou groups. Unconformities or angular unconformities occur at the base of the Horton, Sussex, and Windsor groups. In both the Moncton and Cocagne subbasins, the Windsor–Mabou boundary ranges from conformable in the basin centre to unconformable on its margins. This early basin fill is unconformably overlain by regionally distributed but relatively thin Pennsylvanian–Permian sequences of the Cumberland and Pictou groups.

The Horton Group in the Cocagne Subbasin consists of four formations. From base to top, these comprise 1) coarse redbeds of the Memramcook Formation, 2) red and grey, fine-grained clastic rocks of the McQuade Brook Formation (new name), 3) grey and green fluvial sandstones and mudstones of the Albert Formation, and 4) feldspathic sandstones and red and grey mudstones of the Bloomfield Formation (new name). The unconformably overlying Sussex Group comprises a single thick unit dominated by polymictic conglomerate of the Ridge Brook Formation (new name). The Horton Group in the Moncton Subbasin comprises the same four conformable formations as in the Cocagne Subbasin. In the former subbasin, however, the Albert Formation is characterized by a lacustrine complex of grey strata including organic-rich mudrocks.

The IMDZ is a northeast-trending, structurally complex, fault-bounded panel. It consists of two domains separated by the northeast-striking North River Fault. The northwestern domain is an exhumed Horton section comprising the McQuade Brook Formation, grey lacustrine mudstones and sandstones of the Albert Formation, and green and red sandstones and mudstones of the Indian Mountain Formation (new name). In contrast, the southeastern domain is characterized by exhumed crystalline basement rocks unconformably overlain by the Sussex Group. The Sussex Group in this domain consists of (in ascending order) breccia and conglomerate of the Stilesville Formation (new name), grey and red mudstones of the Gautreau Formation, fine-grained redbeds of the Weldon Formation, and an uppermost coarse-grained redbed sequence, the Briggs Cross Formation (new name).

The Belleisle Fault defines the northwestern boundary of the Cocagne Subbasin with the New Brunswick Platform. Along this fault system, thick proximal basin-margin conglomeratic units are locally preserved in the Horton, Sussex, and Mabou groups. The IMDZ along the southeastern side of the subbasin is interpreted as an inversion zone containing tectonic slices of the Westmorland Uplift, and the Cocagne or Moncton subbasins.

Le sous-bassin de Cocagne dans le Sud-Est du Nouveau-Brunswick est un bassin essentiellement terrestre du Dévonien tardif/Carbonifère précoce à l'intérieur du bassin mixte du Paléozoïque tardif des Maritimes, de l'Est du Canada. On a réalisé une étude stratigraphique et structurale du sous-bassin de Cocagne pour déterminer son lien avec les roches du socle cristallin de délimitation de la plate-forme du Nouveau-Brunswick au nord-ouest et un complexe faillé de strates carbonifères et de strates du socle protérozoïques, la zone déformée du mont Indian, au sud-est. Cette zone déformée est bordée du côté sud-est par le sous-bassin de Moncton et le bombement de Westmorland.

Les composants de remplissage de bassin du Dévonien tardif au Mississippien du sous-bassin de Cocagne sont analogues à ceux du sous-bassin de Moncton et se composent de quatre groupes : le groupe du Dévonien supérieur/Mississippien de Horton, le groupe mississippien de Sussex, Windsor et Mabou. Il existe des discordances stratigraphiques ou angulaires à la base des groupes de Horton, de Sussex et de Windsor. La démarcation entre Windsor et Mabou varie, comme dans le cas du sous-bassin de Moncton, d'une transition concordante dans le centre du bassin à une discordance le long de ses bords. Cette tranche de remplissage de bassin précoce est recouverte sans concordance par des séquences pennsylvaniennes/permiennes des groupes de Cumberland et de Pictou réparties à l'échelle régionale mais relativement minces.

Dans le sous-bassin de Cocagne, le groupe de Horton est constitué, de la base au sommet, de quatre formations : des couches rouges grossières de la Formation de

Memramcook, des roches clastiques à grains fins rouges et grises de la Formation de McQuade Brook (nouveau nom), des mudstones et des grès fluviatiles gris et verts de la Formation d'Albert, et des grès feldspathiques et des mudstones rouges et gris de la Formation de Bloomfield (nouveau nom). Le groupe sus-jacent non concordant de Sussex est constitué d'une seule unité épaisse dans laquelle prédomine un conglomérat polymictique, la Formation de Ridge Brook (nouveau nom). Dans le sous-bassin de Moncton, le groupe de Horton est composé des quatre mêmes formations concordantes que celles présentes dans le sous-bassin de Cocagne, mais il en diffère par le fait que la Formation d'Albert est caractérisée par un complexe lacustre de strates grises comprenant des mudstones riches en éléments organiques.

La zone déformée du mont Indian est un panneau limité par des failles structuralement complexe d'orientation nord-est. Elle est constituée de deux domaines séparés par la faille orientée vers le nord-est de la rivière North. Le domaine nord-ouest correspond à une section exhumée de Horton composée de la Formation de McQuade Brook, de grès et de mudstones lacustres gris de la Formation d'Albert, et de mudstones et de grès verts et rouges de la Formation d'Indian Mountain (nouveau nom). À l'opposé, le domaine sud-est se caractérise par un socle cristallin exhumé, recouvert sans concordance par le groupe de Sussex, qui est composé d'un conglomérat et d'une brèche basaux, de la Formation de Stilesville (nouveau nom), des mudstones sus-jacents gris et rouges de la Formation de Gautreau, des couches rouges à grains fins subséquentes de la Formation de Weldon, et d'une séquence supérieure de couches rouges à gros grains, la Formation de Briggs Cross (nouveau nom).

La ligne de démarcation nord-ouest du sous-bassin de Cocagne avec la plate-forme du Nouveau-Brunswick est définie par la faille de Belleisle. D'épaisses unités conglomératiques proximales de bordure de bassin sont localement préservées dans les groupes de Horton, de Sussex et de Mabou le long de cette faille. Le côté sud-est du sous-bassin, la zone déformée du mont Indian, est interprété en tant que zone d'inversion renfermant des tranches tectoniques du bombement de Westmorland ainsi que des sous-bassins de Cocagne et de Moncton.

INTRODUCTION

Over the past six decades, a number of regional stratigraphic studies have been undertaken in the Upper Devonian–Carboniferous Maritimes Basin of southeastern New Brunswick (Fig. 1), the most comprehensive of which is Gussow's seminal work (Gussow 1953). Greiner (1962), Shroder (1963), Popper (1965), McCutcheon (1981), Carter and Pickerill (1985a), Browne (1990), and St. Peter (1992) have also provided papers on specific aspects of Carboniferous stratigraphy in the province.

Within the same time period, several studies of key structural elements have been conducted that are pertinent to interpreting the sedimentation and deformation history of

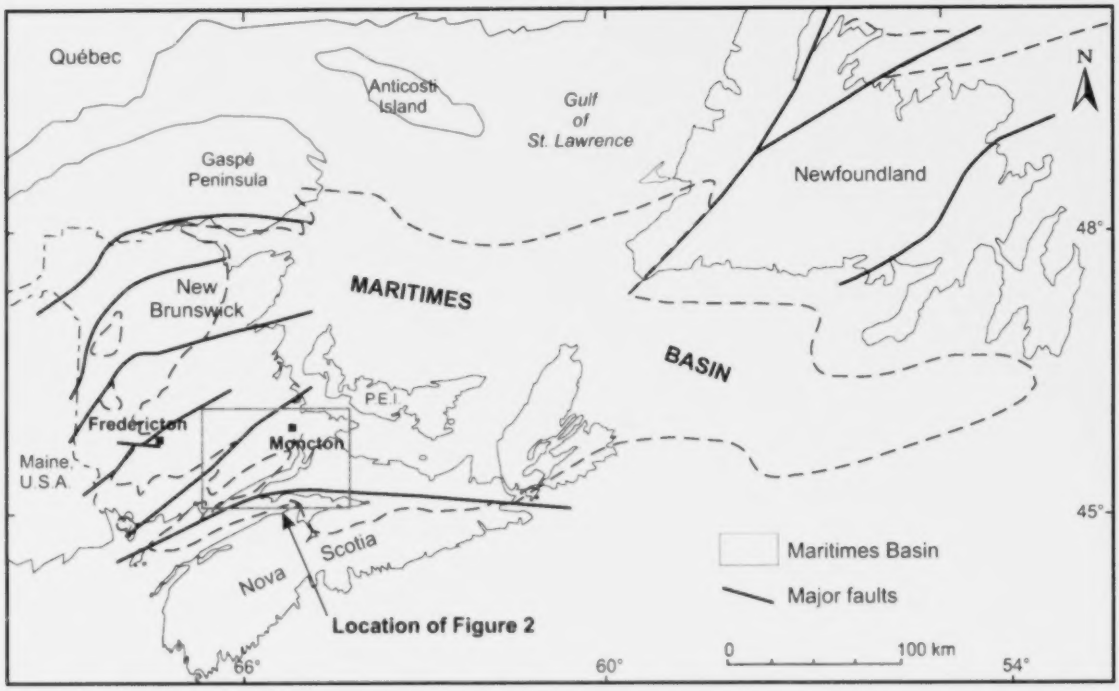


Figure 1. Distribution of the Upper Devonian–Permian rocks of the Maritimes Basin, eastern Canada.

stratigraphic sequences during the four tectonic events (manifested as unconformities) that punctuate the Carboniferous section. The relevant structural papers include Gussow (1953), Webb (1963), Belt (1968), Ruitenberg and McCutcheon (1982), Leger and Williams (1986), St. Peter (1993), Nickerson (1994), and Park and St. Peter (2005). Additional stratigraphic and structural details are available from many bedrock maps published by the Geological Survey of Canada and the New Brunswick Department of Natural Resources (DNR).

A synthesis of the foregoing stratigraphic and structural work reveals that various conflicts or debates exist among authors concerning 1) the position and correlation of some lithologic units within the stratigraphic column, and 2) the location, kinematics, and timing of major regional faults. The present author has conducted detailed field investigations over several summer seasons, seeking evidence that might resolve some of the interpretive differences. This report summarizes the existing and new evidence, the latter of which was gained particularly from the Indian Mountain Deformed Zone (IMDZ) and Cocagne Subbasin (Fig. 2).

The present report gives new evidence that relates to the contact relationships (such as unconformities) between groups, and that concerns the assignment or reassignment of lithostratigraphic units. The new evidence is used to re-examine the stratigraphy and structure of the Cocagne and Moncton subbasins and the IMDZ, and attempts to resolve how the stratigraphic section in the IMDZ is related to its bounding subbasins. The report also describes the stratigraphic contrasts between subbasins and their adjacent uplifts, and summarizes major and minor faults that influenced the inception and/or inversion of the Cocagne Subbasin and northwestern Moncton Subbasin.

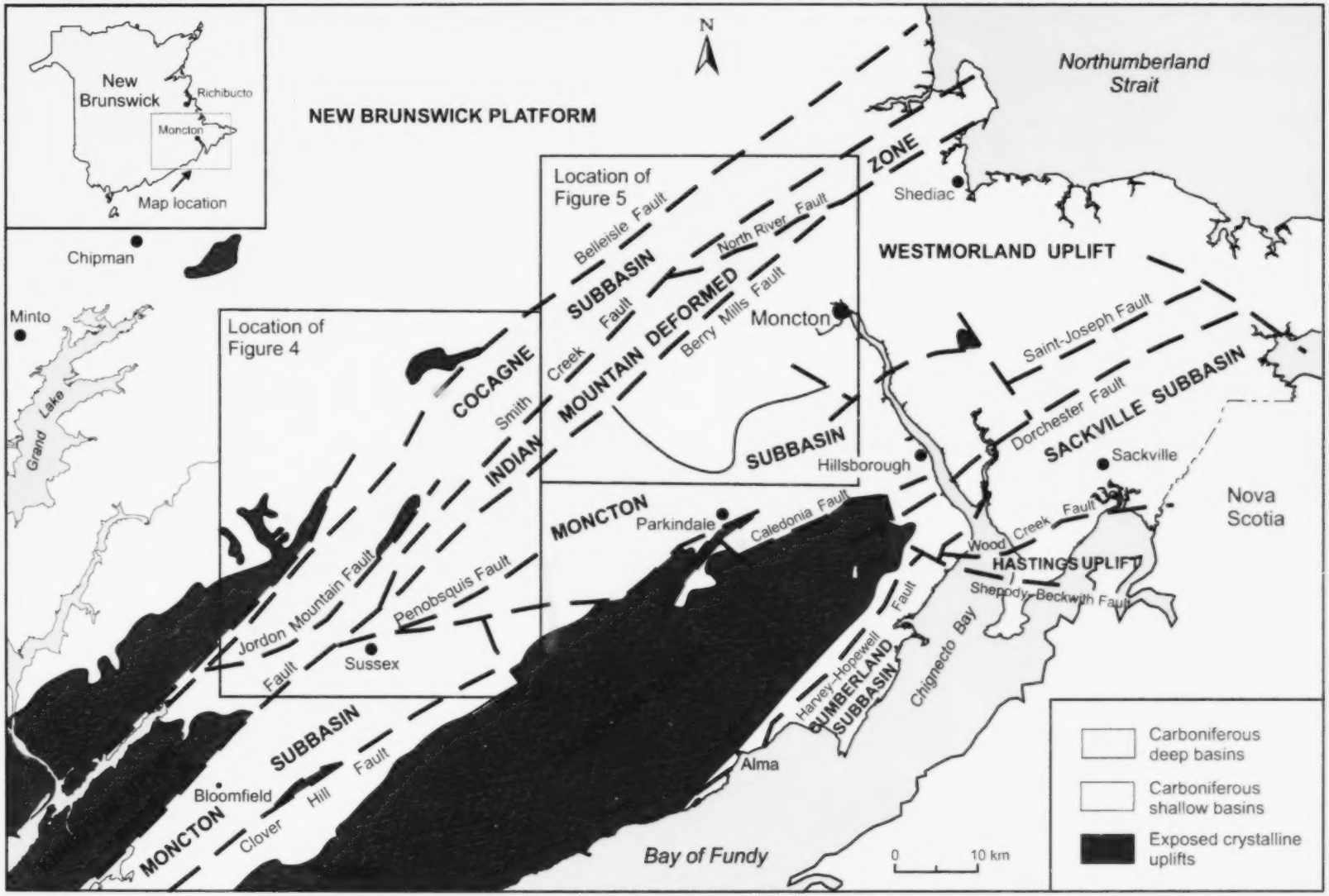


Figure 2. Distribution of uplifts and subbasins of the Maritimes Basin in southeastern New Brunswick.

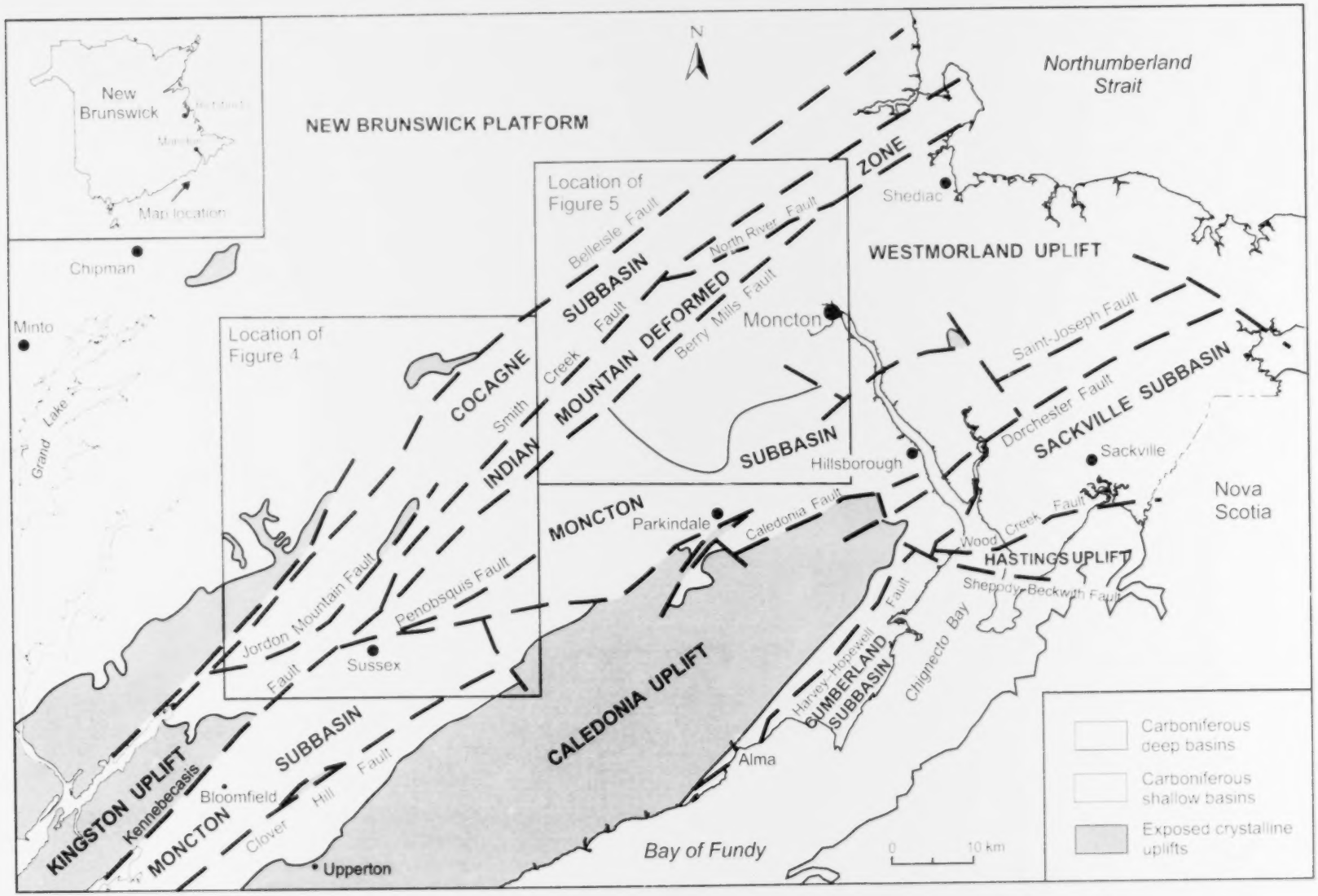


Figure 2. Distribution of uplifts and subbasins of the Maritimes Basin in southeastern New Brunswick.

REGIONAL GEOLOGY

Maritimes Basin

The Upper Devonian–Permian rocks of New Brunswick are part of the largely terrestrial Maritimes Basin of eastern Canada (Fig. 1), which formed as a post-Acadian successor basin (Roliff 1962). The Maritimes Basin overlies the Proterozoic and early Paleozoic Appalachian Orogen (Hibbard et al. 2006). It covers an area of about 150 000 km^2, about 70% of which lies offshore in the Gulf of St. Lawrence and beneath the continental shelf.

The Maritimes Basin is a composite structure comprising a series of depocentres separated by intervening horsts or uplifts (Fralick and Schenk 1981; Bradley 1982; Hamblin 1989, 1992; Durling and Marillier 1990; St. Peter 1993). The component depocentres in southeastern New Brunswick are the Cumberland, Sackville, Moncton, and Cocagne subbasins (Fig. 2). These subbasins are extensional or transtensional depocentres and are examples of the early phase of Maritimes Basin evolution (Bradley 1982). Thinner or condensed sections lie adjacent to the subbasins on the margins or 'shoulders' of the horsts or uplifts, those being the Caledonia Uplift, Westmorland Uplift, and New Brunswick Platform (Fig. 2).

The Maritimes Basin fill in New Brunswick is regionally divisible into two parts, separated by a basin-wide middle Carboniferous (Namurian) unconformity (MacIntosh and St. Peter 2005).

1. Below the unconformity (Fig. 3), the fill comprises a basal alluvial-lacustrine succession (Horton Group) overlain in turn by a terrestrial redbed clastic succession (Sussex Group), a mainly marine mixed clastic-carbonate-evaporite sequence (Windsor Group), and another terrestrial redbed succession (Mabou Group).

   These Upper Devonian–Mississippian strata of the Horton to Mabou groups are interpreted to fill a complex system of subbasins that collectively evolved in a transtensional-transpressional setting. Unconformities between the Horton and Sussex groups and between the Windsor and Mabou groups may reflect alternating tensional and compressional regimes. The Upper Devonian–Mississippian stratigraphic section in each subbasin is broadly similar, suggesting that the depocentres underwent comparable tectonic and perhaps paleoclimatic histories (MacIntosh and St. Peter 2005).

2. Above the Namurian unconformity (Fig. 3), fluvial-floodplain-mire successions of the Lower Pennsylvanian Cumberland Group and Upper Pennsylvanian–Permian Pictou Group are interpreted to have infilled a thermal subsidence or sag basin (Bradley 1982). This Pennsylvanian–Permian section forms a more regional cover sequence that

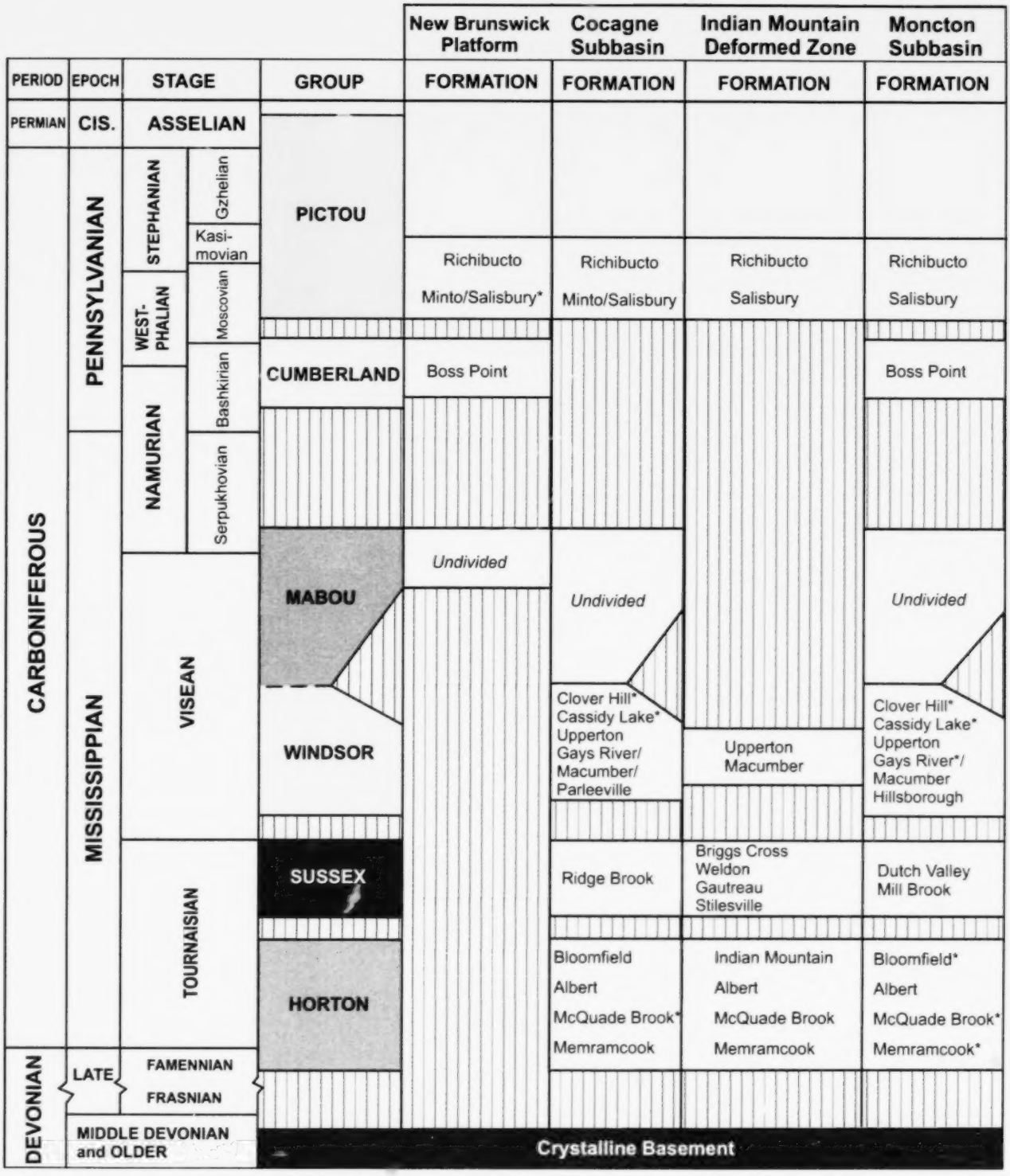


Figure 3. Stratigraphic columns of Upper Devonian–Carboniferous rocks for the New Brunswick Platform, Cocagne Subbasin, Indian Mountain Deformed Zone, and Moncton Subbasin, Maritimes Basin, southeastern New Brunswick. * indicates that the formation is either present only in the subsurface of that zone or is exposed beyond the report area shown on Figures 4 and 5.

unconformably overlies the older subbasins and partly to completely buries the adjoining uplifts. The Westmorland Uplift, in particular, is a largely buried basement block that lies south of the Berry Mills Fault and north of the Moncton Subbasin (Fig. 2). It is characterized by the absence (or near absence) of Horton strata and the presence of relatively thin cover sequences of younger Carboniferous groups.

Cocagne Subbasin

The Cocagne Subbasin is a structural subbasin containing the erosional remains of a depocentre (or depocentres) that is (are) characterized by distinctive stratigraphy in the Horton, Sussex, and Windsor groups (Fig. 3). The northwestern boundary of the Cocagne Subbasin is the contact with the New Brunswick Platform along the Belleisle Fault and is interpreted to approximate the depositional margin of the basin (Fig. 2).

The southeastern boundary is less clearly defined but, for the purpose of stratigraphic description, is delineated here by the Smith Creek and Kennebecasis faults (Fig. 2, 4, 5). This delineation is not meant to imply, however, that either fault represented a southeastern depositional boundary of the Cocagne Subbasin during sedimentary cycles of the Horton, Sussex, or Windsor groups.

Indian Mountain Deformed Zone

The Indian Mountain Deformed Zone (IMDZ) is a northeast-trending, structurally complex panel up to 7 km wide, bounded by the Smith Creek Fault on the northwest and the Berry Mills Fault on the southeast (Fig. 2, 4, 5). It forms a structural inversion zone that largely separates the Cocagne Subbasin on its northwestern margin from the Moncton Subbasin on its southeastern margin (Fig. 2). The zone consists of two domains divided by the northeast-striking North River Fault (Fig. 6). The northwestern domain is an exhumed section of the Horton Group. In contrast, the southeastern domain is characterized by exhumed crystalline basement rocks with minor erosional outliers of the Horton Group, unconformably overlain by distinctive rocks of the Sussex Group.

Moncton Subbasin

The Moncton Subbasin is distinguished from the Cocagne Subbasin and the IMDZ on the basis of its distinctive stratigraphy in the Horton and Sussex groups, and the presence of a coarse terrestrial clastic unit (Hillsborough Formation) at the base of the predominantly marine Windsor Group (Fig. 3). The southeastern margin of the Moncton Subbasin is defined as crystalline rocks of the exhumed or buried Caledonia Uplift (Fig. 2). For the purpose of stratigraphic description, the northwestern boundary is delineated here as the Berry Mills and Kennebecasis faults (Fig. 2, 4, 5).

STRATIGRAPHY

The regional lithostratigraphy within the Maritimes Basin has been summarized by Roliff (1962), Kelley (1967), Hacquebard (1972), Williams (1974), Gibling (1995), and van de Poll (1995). In the past, the stratigraphic section has been divided into five lithostratigraphic groups which, in ascending order, are the Horton, Windsor, Mabou, Cumberland, and Pictou groups (Ryan et al. 1991). After St. Peter (2002) identified an unconformity in the Horton Group in New Brunswick, beds above the unconformity (but beneath the Windsor Group) were assigned to the Sussex Group (Fig. 3).

Basement Rocks

Basement rocks underlying the Maritimes Basin in southern New Brunswick range from Proterozoic gneiss to Middle Devonian plutonic, volcanic, and sedimentary rocks. Williams (1978, 1979) and Williams and Hatcher (1982) divided the Ordovician and older basement rocks of the northeastern Appalachians into five tectonostratigraphic zones. From northwest to southeast, these are the Humber, Dunnage, Gander, Avalon, and Meguma zones. The Maritimes Basin in southern New Brunswick overlies the Avalon and Gander zones or terranes.

The Avalon–Gander terrane boundary is not a simple, through-going crustal fault (see Whalen et al. 1996). Its location is considered to be a zone approximately coincident with the present position of the Taylor Brook Fault (Fig. 4), which marks the suture between Cambro–Ordovician sedimentary rocks of the St. Croix Belt on the north and Silurian volcanic and sedimentary rocks of the Mascarene Belt on the south (Johnson 2003; Thorne et al. 2005).

Northeast of the Taylor Brook break, the location of the Avalon–Gander terrane boundary beneath the Carboniferous is uncertain. However, if the Taylor Brook Fault links with the Belleisle Fault northeast of Harry Brook (Fig. 4), then the long-term mobility apparently associated with this zone almost certainly had a controlling influence on the northwestern boundary of the Cocagne Subbasin (Fig. 4, 5).

Basement rocks are exposed along the northwestern margin of the Cocagne Subbasin in the Marrtown and Salem areas and as inliers within the New Brunswick Platform (Fig. 2, 4). An inlier of Silurian volcanic and sedimentary rocks occurs within the Cocagne Subbasin at Jordan Mountain. As well, a much-faulted Proterozoic complex of cataclastic granite, diorite and minor amphibolite, schist, and marble is situated at The Gorge north of Moncton in the IMDZ (Fig. 6, 7; St. Peter and Park 2006a).

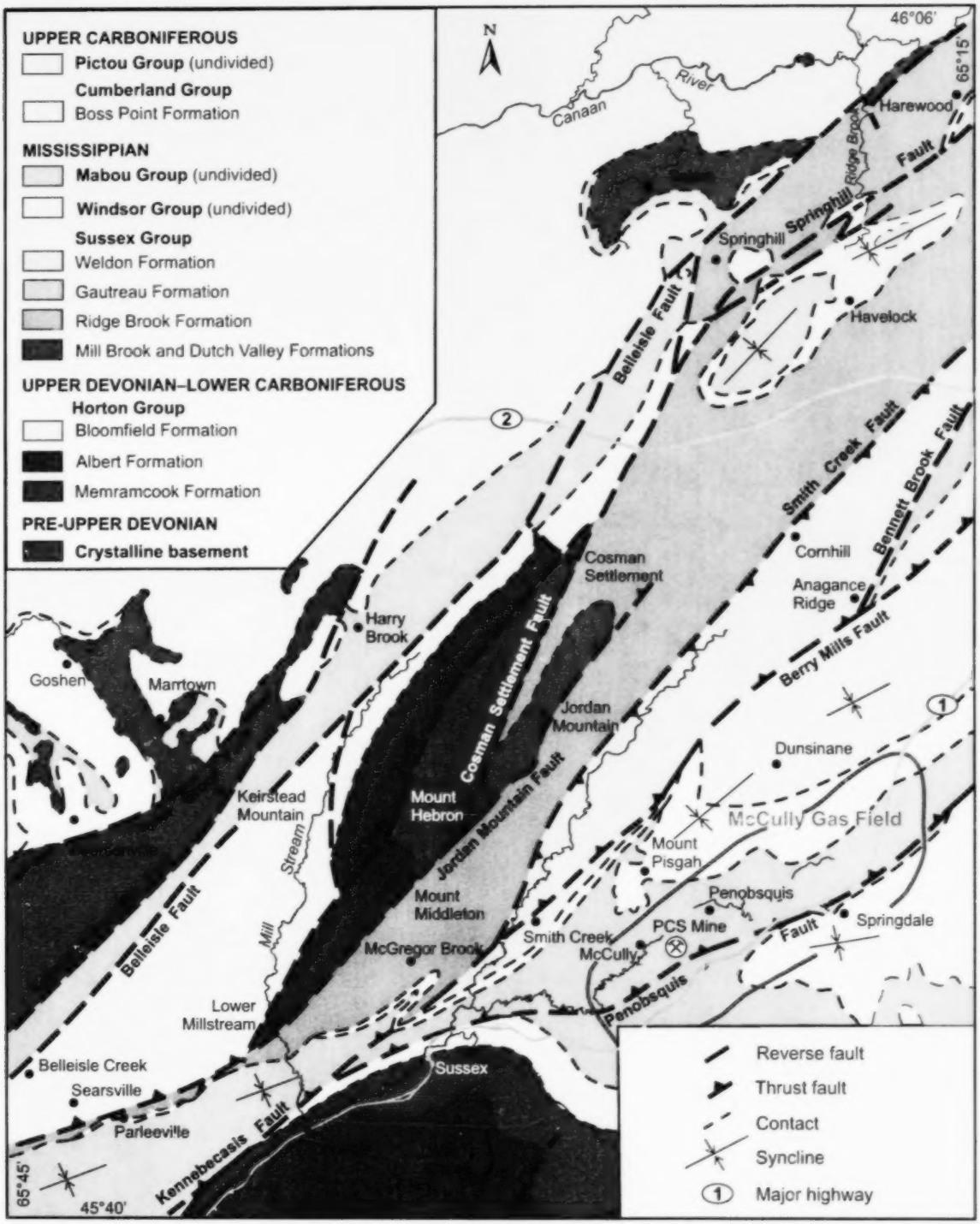


Figure 4. Simplified geological map of the southwestern part of the Cocagne Subbasin and Indian Mountain Deformed Zone of the Maritimes Basin, southeastern New Brunswick. See Figure 2 for map location.

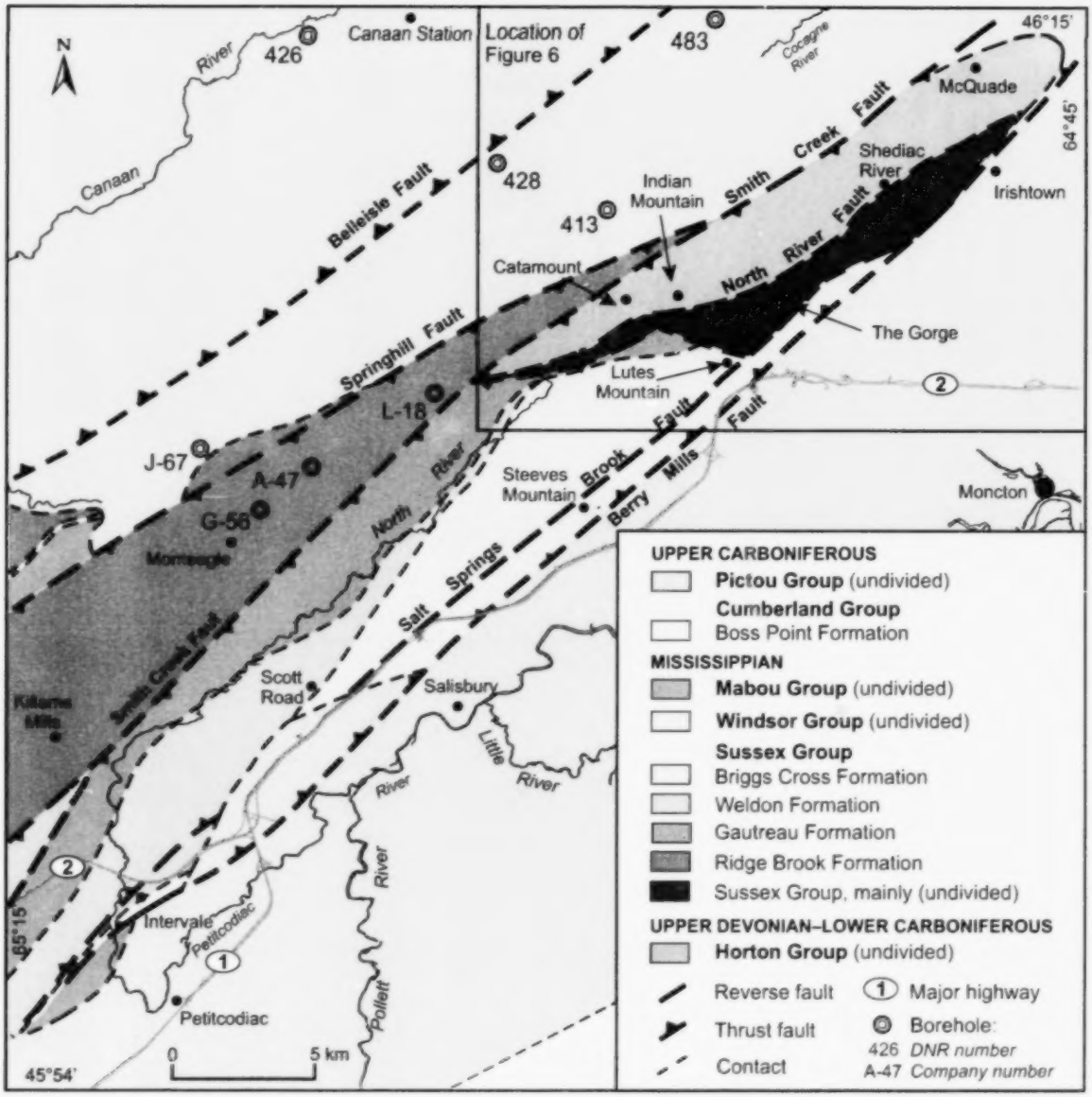
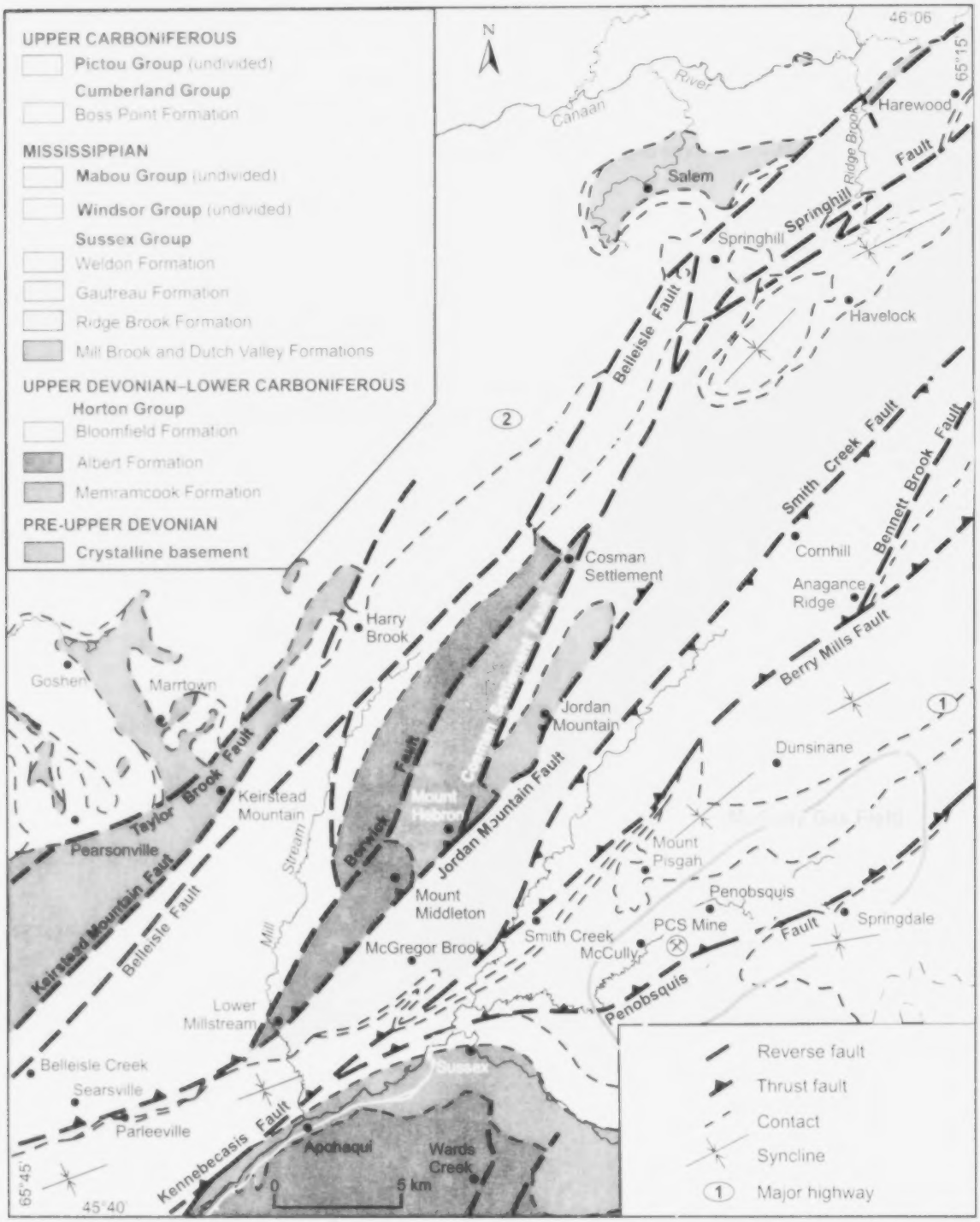


Figure 5. Simplified geological map of the northeastern part of the Cocagne Subbasin and Indian Mountain Deformed Zone of the Maritimes Basin, southeastern New Brunswick. See Figure 2 for map location.

Horton Group

Rocks of the Horton Group range in age from Late Devonian to Early Mississippian (early Tournaisian) and represent the oldest strata of the Maritimes Basin in this part of New Brunswick. The stratigraphy of this group varies between the Cocagne Subbasin, IMDZ, and Moncton Subbasin (Fig. 4, 5, 6). In the Cocagne Subbasin, the Horton Group comprises (in ascending order) the Memramcook, McQuade Brook, Albert, and Bloomfield formations.



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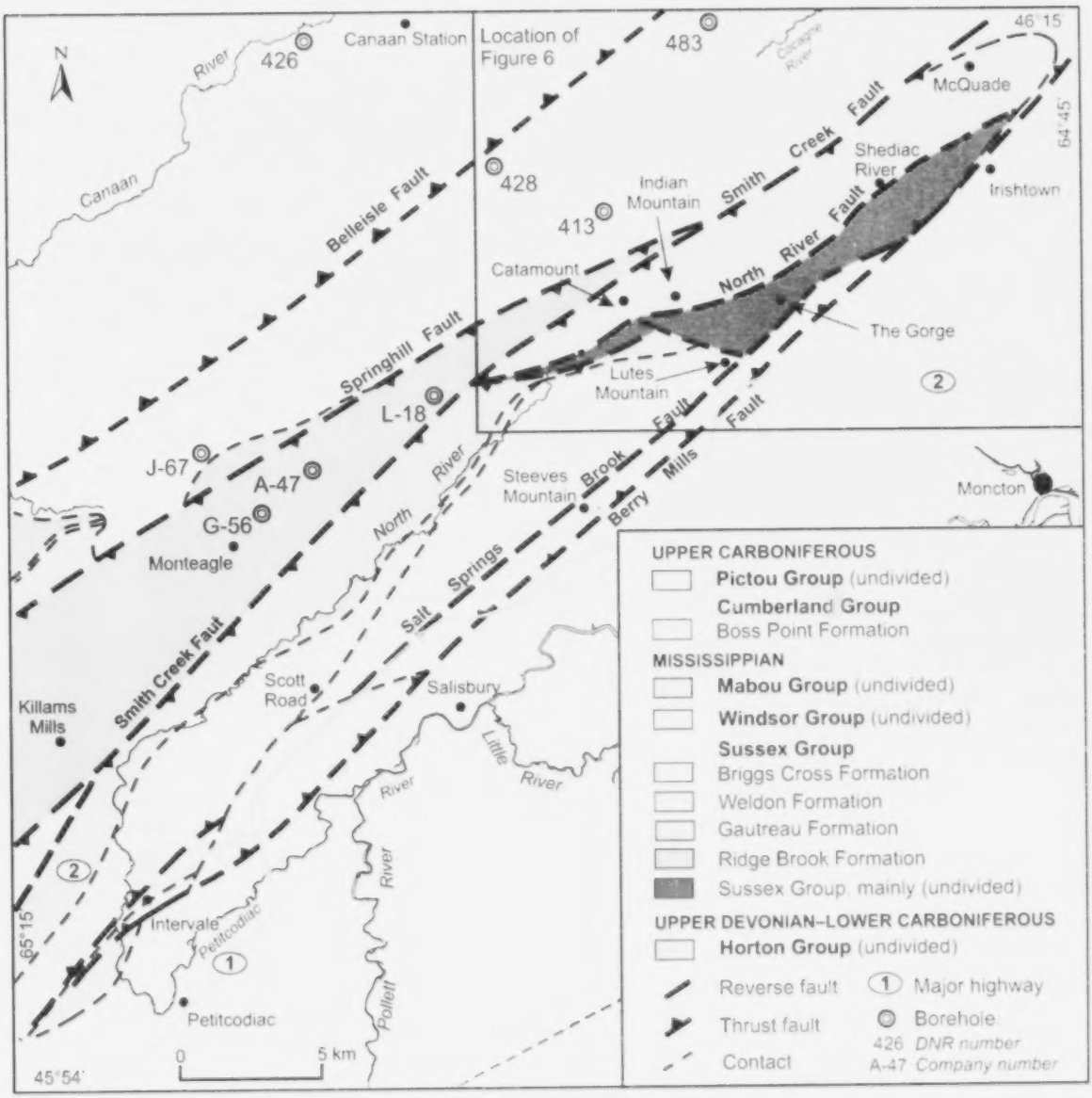

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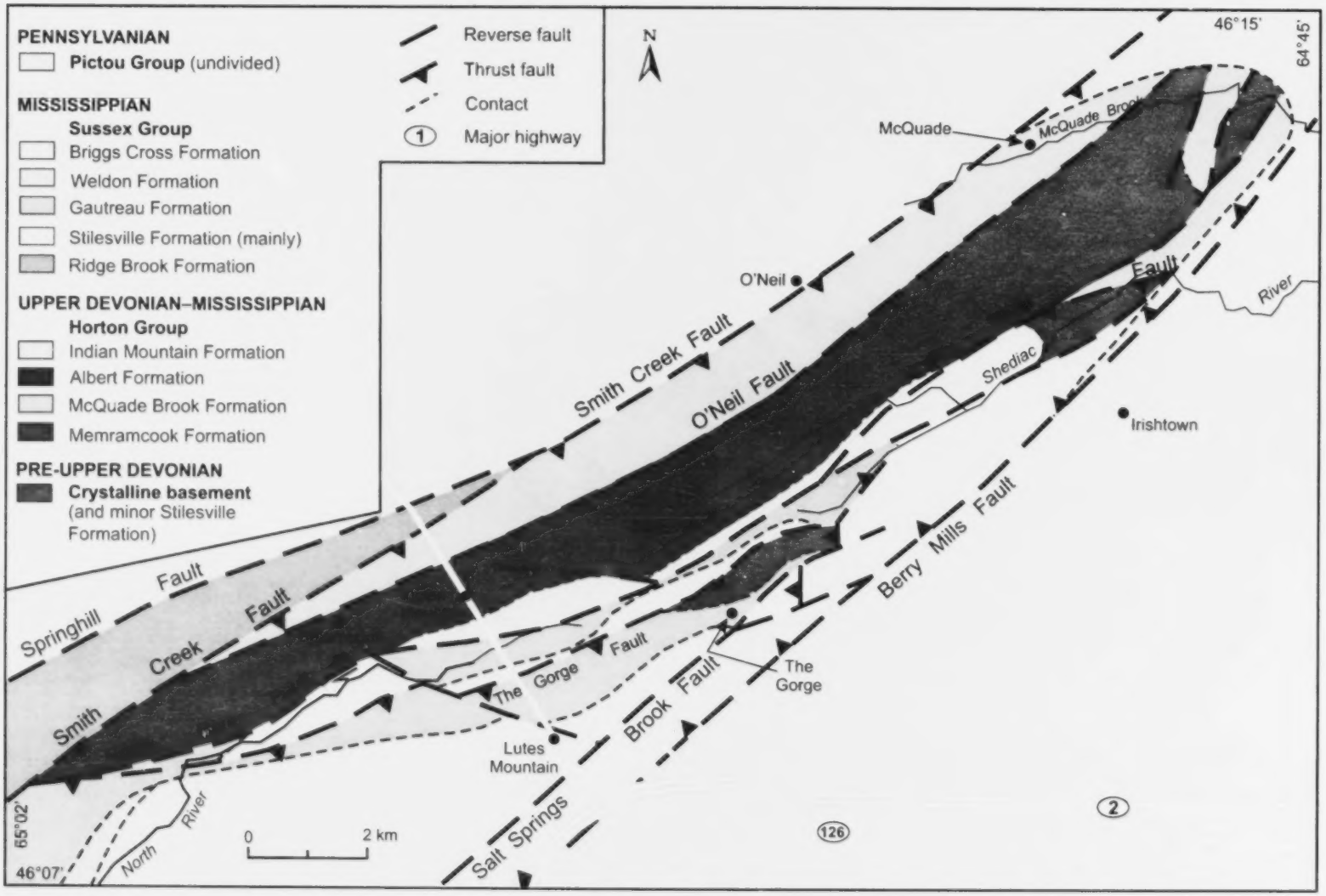

Figure 6. Geological map of the Indian Mountain and McQuade Brook area (after St. Peter and Park 2006b, 2006c). See Figures 2 and 5 for map location.

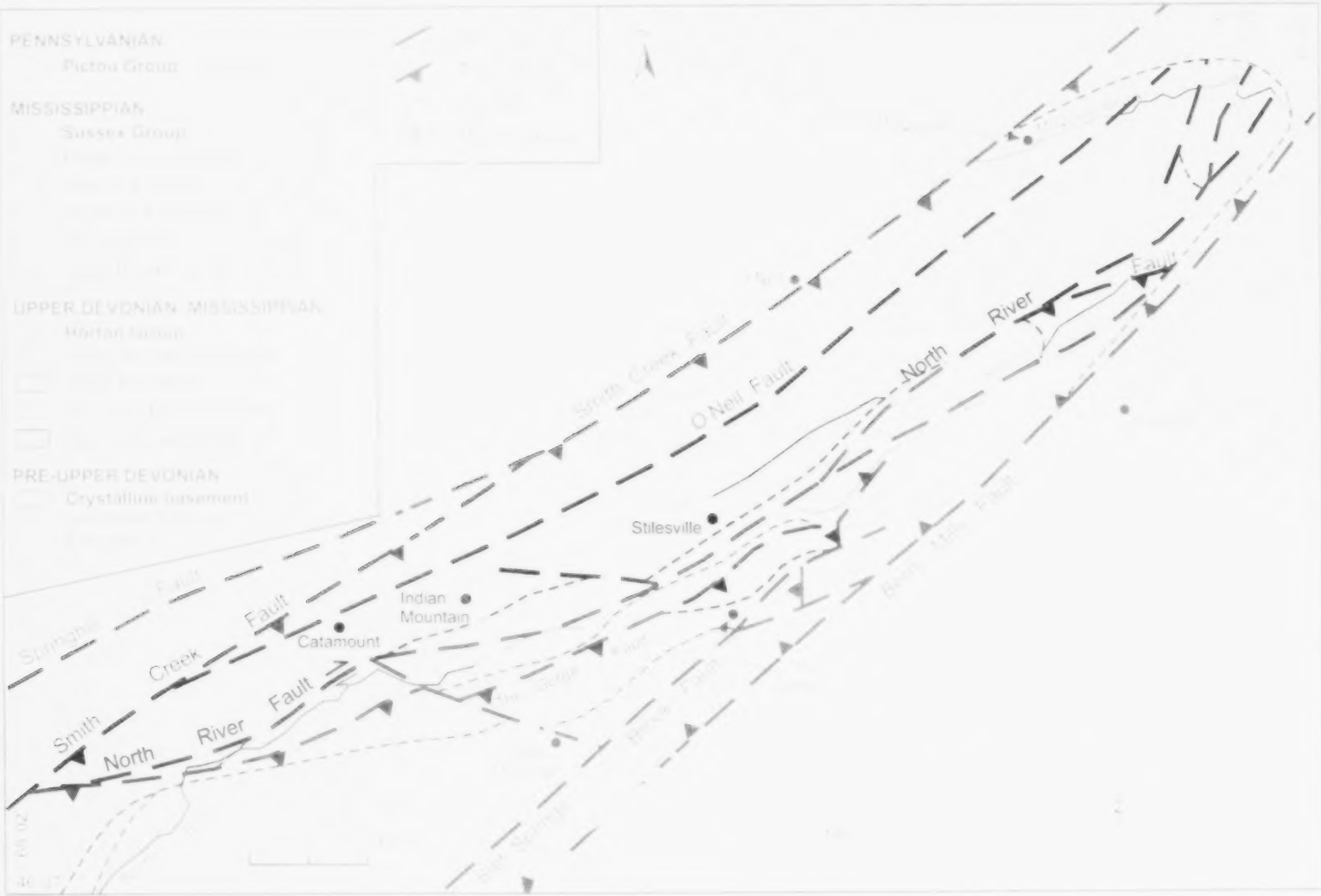
PENNSYLVANIAN
Pictou Group
MISSISSIPPIAN
Sussex Group
UPPER DEVONIAN-MISSISSIPPIAN
Horton Group
PRE-UPPER DEVONIAN
Crystalline basement
Stilesville
Indian Mountain
Catamount
Smith Creek Fault
O'Neil Fault
North River Fault
Smith Creek Fault
North River Fault

Figure 6

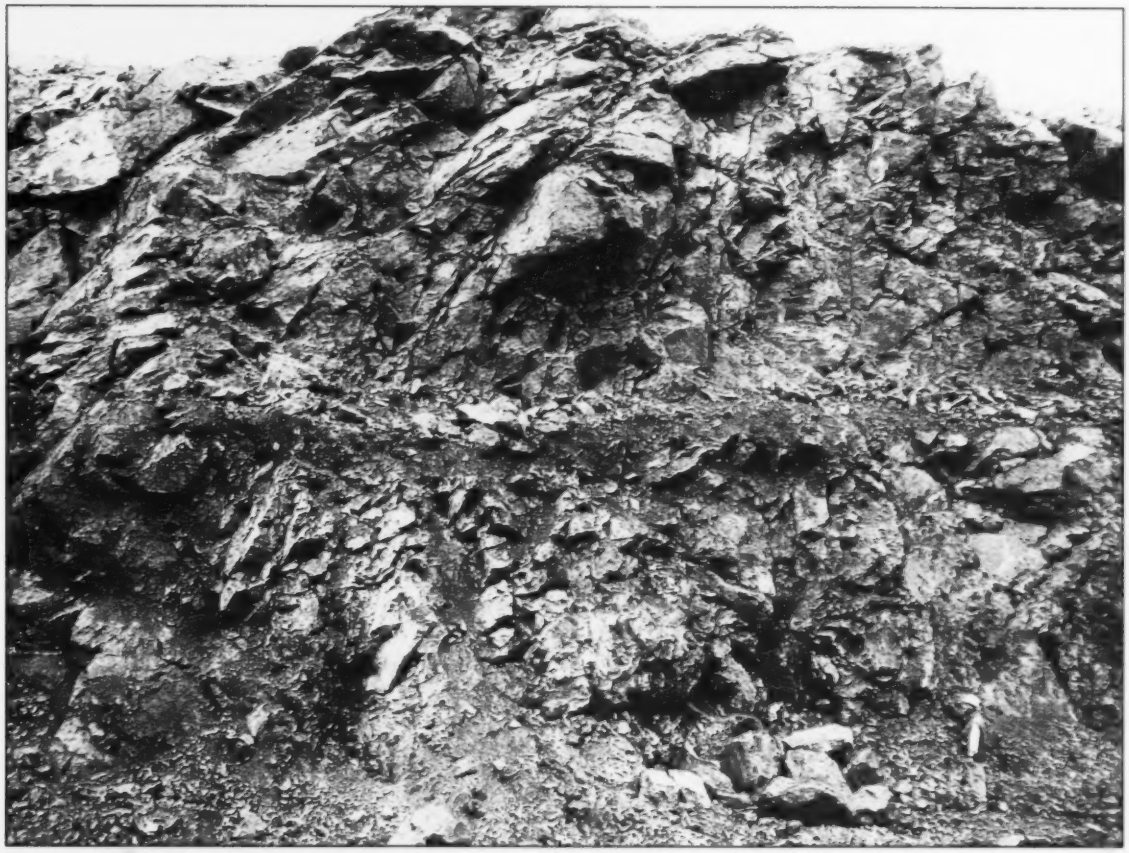

Figure 7. Dark grey diorite intruded by a fine-grained pink granite dyke, which traverses from lower centre to upper left. The dyke and host diorite are both highly fractured. The exposure is situated in the Tri-Gil Quarry at The Gorge (Fig. 6). Person in lower right for scale.

In the northwestern domain of the IMDZ, the group consists of the Memramcook, McQuade Brook, Albert, and Indian Mountain formations. In the eastern Moncton Subbasin, the Horton section is represented by the Memramcook, McQuade Brook, and Albert formations, whereas in the western Moncton Subbasin, it comprises the Memramcook, Albert, and Bloomfield formations (Fig. 3).

Memramcook Formation

The type locality of the Memramcook Formation occurs in the Moncton Subbasin along Stony Creek about 15 km southeast of Moncton (Fig. 2; Norman 1941a). There, the formation comprises repetitively stacked, fining-upward sequences of red and grey polymictic conglomerate, sandstone, and mudstone (McLeod and Ruitenberg 1978; Carter and Pickerill 1985a; St. Peter 1992). However, the Memramcook Formation is not exposed in the northwestern part of the Moncton Subbasin being discussed here (Fig. 4, 5).

In the Cocagne Subbasin on the west side of Jordan Mountain, the Memramcook Formation is exposed in a faulted wedge reaching from Mount Hebron in the south to Cosman Settlement in the north (Fig. 4). The formation in the Jordan Mountain area is dominated by

pinkish red to reddish brown to rarely grey, angular- to rounded-clast, polymictic granule to boulder conglomerate; massive to flat- and cross-bedded feldspathic sandstones; and minor siltstone and mudstone. The sandstones contain abundant pink and white feldspar and, if point-counted, would probably plot mainly in the arkosic field. The conglomerate carries abundant granitic and lesser metasedimentary, mafic, and felsic volcanic clasts. On the southwest end of the Jordan Mountain basement block, beds of the Memramcook Formation strike northwest and dip southwest off the crystalline rocks with apparent unconformity. Southwest of Jordan Mountain, in the Mount Middleton area, the formation is gradationally overlain by the Albert Formation.

In the Monteagle area of the Cocagne Subbasin (Fig. 5), the Memramcook Formation has been intersected in oil and gas exploration boreholes A-47 and L-18 drilled by Columbia Natural Resources Canada Ltd. (hereafter referred to as Columbia). The formation in these boreholes is dominantly red, coarse, polymictic conglomerate with minor interbeds of lithic sandstones and mudstones, and has been spore-dated (see below). In the nearby borehole G-56, an analogous but undated conglomerate-dominated section intersected between 1259 m and 1500 m (TD) below surface may represent Memramcook strata.

At borehole J-67 northwest of borehole G-56 (Fig. 5), a section of massive conglomerates was intersected below the Minto Formation from a depth of 14 m below surface to a total depth of 2499 m. That section contains rare red mudstone intervals with one thin grey mudstone bed at 2025 m. The grey mudstone was barren of spores (Dolby 2004), and a cursory core examination indicated no obvious formational division within the massive conglomerates. The conglomerates, at least in the lower part of the borehole, presumably represent the Memramcook Formation.

In the IMDZ, the Memramcook Formation outcrops along Shediac River on the south side of the North River Fault about 1.5 km north of Irishtown (Fig. 6). Memramcook beds in that location are arranged into fining-upward sequences of pinkish red to grey, angular-clast, granule to pebble and cobble conglomerate; reddish brown to buff and rare grey or green, medium-grained to pebbly, cross-bedded and massive lithic sandstone; and lesser beds of reddish brown to grey or olive, fine-grained sandstone, siltstone, and mudstone (Fig. 8, 9). The finer grained redbeds locally have carbonate nodules, and the finer grained grey beds in places contain plant detritus.

The Memramcook Formation along Shediac River is interpreted to be in reverse- or thrust-faulted contact with the McQuade Brook Formation along the North River Fault (Fig. 6). South of there, Memramcook rocks are in fault contact with those of the Briggs Cross Formation along a northeast-striking splay of the North River Fault. To the west, the Memramcook Formation is believed to be unconformably succeeded by basal conglomerates of the Stilesville Formation (Sussex Group).

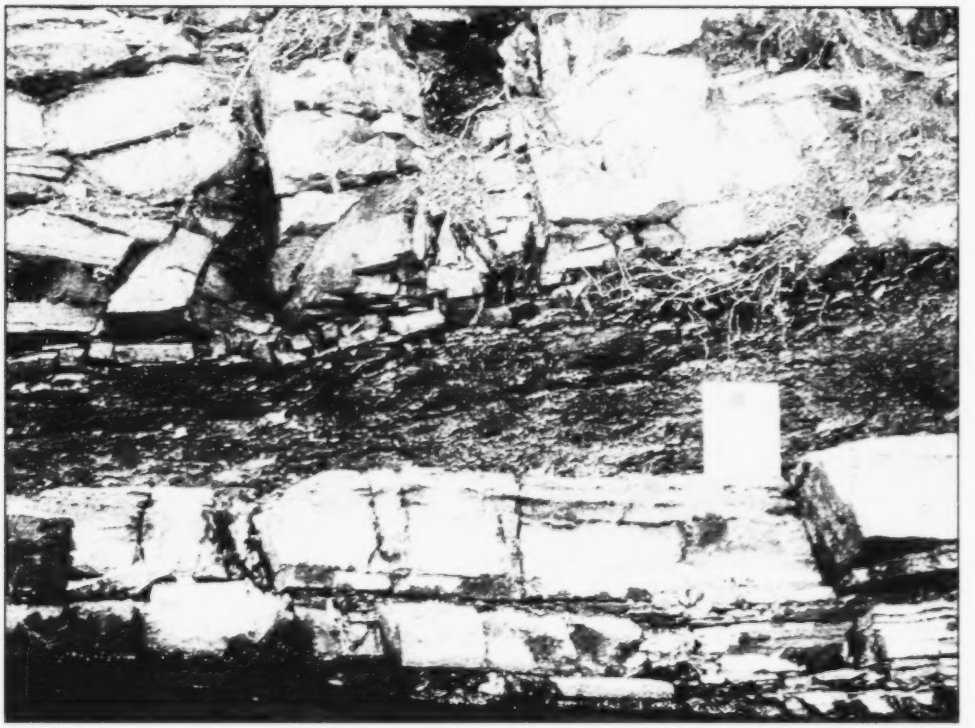

Figure 8. Exposure of the Memramcook Formation at Shediac River, showing cycles of buff sandstone and reddish brown mudstones. Clipboard is 23 cm wide. (Note: green coloration of the outcrop surface is due to a mossy overgrowth.)

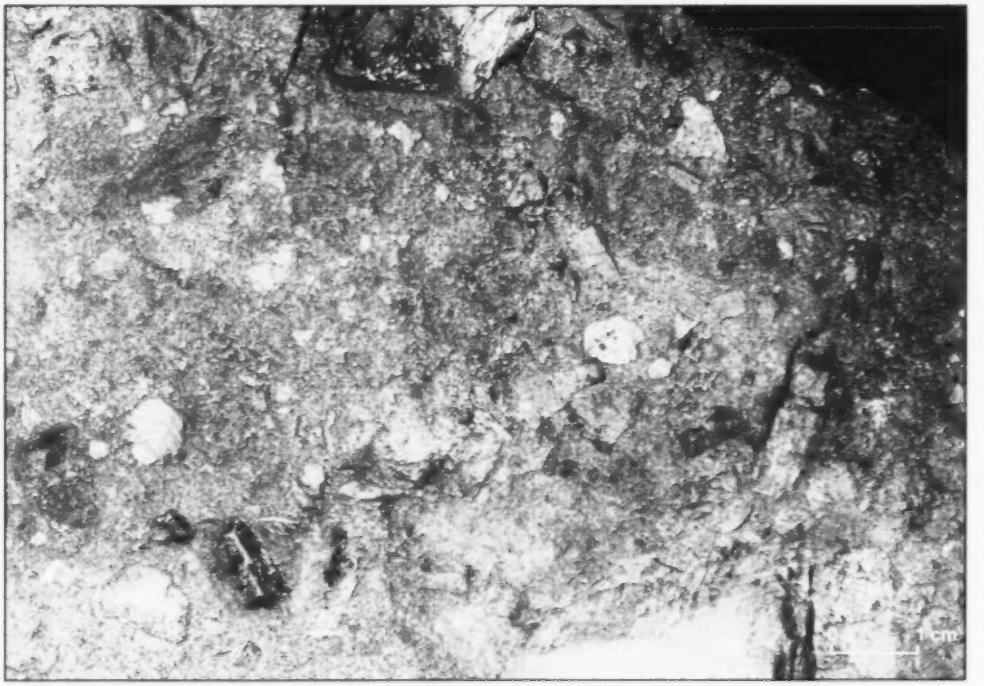

Figure 9. Conglomerate from the base of a fining-upward cycle in the Memramcook Formation at Shediac River. Most of the clasts are locally derived angular pink granite.

Three spore-bearing samples were recovered from rare grey mudstones within coarse redbeds of the Memramcook Formation in two boreholes (A-47 and L-18) northeast of Monteagle (Fig. 5). The formation there was dated at Late Devonian (Famennian) by Dolby (2001a, 2004). No grey siltstone or mudstone beds that potentially could yield palynomorphs have been found in the Memramcook strata near Jordan Mountain. Spore assemblages from seven samples collected along Shediac River in the IMDZ indicated a Tournaisian early Zone 1 to possibly Zone 2A age (Fig. 10) for the Memramcook Formation (Dolby 1991, 1995a, 2002, 2004, 2006a).

McQuade Brook Formation

McQuade Brook Formation is the new name for a unit of rocks situated in the northwestern domain of the IMDZ. The beds in the IMDZ herein assigned to the McQuade Brook Formation previously were mapped as the Memramcook Formation by Norman (1941b) and Gussow (1953).

The best and most continuous section of the McQuade Brook Formation occurs along McQuade Brook. About 1.5 km south of the brook, the Albert Formation is interpreted to concordantly overlie the McQuade Brook Formation, although the contact between the two units is not exposed in that location. Good exposures of the newly named formation occur along small creeks between the O'Neil and Smith Creek faults from Catamount to McQuade Brook (Fig. 6). Beds in this area dip southeast and are in contact with the Albert Formation along the O'Neil Fault. The McQuade Brook Formation is also visible along Shediac River on the north side of the North River Fault.

At McQuade Brook and elsewhere in the IMDZ, the formation comprises reddish brown or grey mudstone and siltstone interbedded with reddish brown or grey, very fine- to fine-grained, parallel- and ripple-laminated sandstone; and minor grey to reddish grey, medium- to coarse-grained, flat- or cross-bedded sandstone (Fig. 11). Exposures along Shediac River adjacent to the North River Fault are mainly a tectonic breccia of sandstone, siltstone, mudstone, and shale. Grey to reddish grey, granular to small pebble conglomerates with intraclasts of mudstone and/or micritic limestone are seen locally. Thus, the McQuade Brook Formation comprises primarily fine-grained clastic rocks in contrast with the Memramcook Formation, which consists of predominantly coarse-grained clastic rocks.

Fine-grained sandstones and mudstones that are lithologically and stratigraphically analogous to the McQuade Brook Formation are exposed in the Memramcook area of the Moncton Subbasin (Fig. 2). These beds previously were assigned to either the upper part of the Memramcook Formation (Norman 1941a; Gussow 1953; McLeod and Ruitenberg 1978) or the basal part of the Albert Formation (St. Peter 1992). St. Peter (1992) mapped these strata in the subsurface using borehole data and seismic reflection profiles from

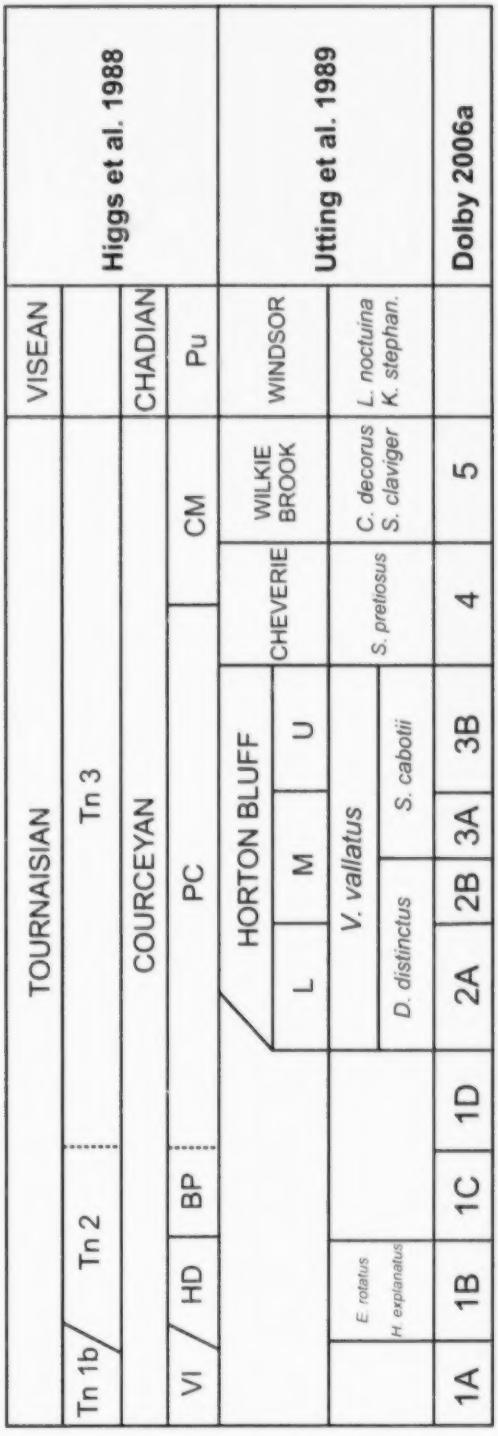

Figure 10. Lower Carboniferous (Tournaisian) Spore Zonation Chart (modified from Utting et al. 1989 and Dolby 2006a).

Figure 11. McQuade Brook Formation at type section along McQuade Brook, showing a thick bed of grey, fine-grained sandstone overlain by a very thick interval of reddish brown mudstone. Hammer for scale.

Memramcook westward to Parkindale (Fig. 2). St. Peter and Johnson (1997) reassigned the beds to the McQuade Brook Formation.

The McQuade Brook Formation has been recognized in two of the four Columbia boreholes near Monteagle (Fig. 5). In borehole G-56, a faulted and partly overturned sequence of fining-upward cycles of mainly red sandstone, siltstone, and mudstone occurs between the depths of 642 m and 1259 m below surface and concordantly overlies the Memramcook Formation. The sequence includes minor beds of grey sandstone and mudstone and, in the upper part of the section, rare beds of black shale. In borehole L-18, very similar fining-upward cycles of sandstone and mudstone with rare beds of conglomerate occur at a depth of 260 m to 548 m below surface. The generally fine-grained red and grey strata that characterize the McQuade Brook Formation are not recognized in sections of the Horton Group in the southwestern part of either the Cocagne or Moncton subbasins.

Palynomorph assemblages have been collected from seven outcrops of the McQuade Brook Formation in the IMDZ. Five assemblages provided only a general Tournaisian or Tournaisian Zone 1 age, but the two other assemblages yielded a more constrained Zone 1C to 1D age (Fig. 10; Dolby 1995a, 2004). Three samples from Columbia boreholes G-56 and L-18 in the Monteagle area of the Cocagne Subbasin (Fig. 5) indicated a Tournaisian Zone 1A age for beds near the base of the formation and a Zone 2 or even Zone 2B age for stratigraphically higher beds. It is possible that the younger, Zone 2B assemblage may derive from a fault-bounded interval of Albert Formation in borehole L-18.

Albert Formation

Norman (1941a) first introduced the name Albert Formation to describe the grey beds in the medial part of the Horton Group. Gussow (1953) used Norman's definition and mapped the Albert Formation throughout the Late Devonian–Carboniferous basins of southeastern New Brunswick. Gussow's (1953) definition and distribution of the Albert Formation was generally followed by subsequent workers until 2003, when St. Peter collected and had analyzed a spore assemblage from grey shales exposed 25 km south of Moncton along Peck Creek. Although those beds had been assigned to the Albert Formation (McLeod 1980) or Weldon Formation (Norman 1941a; St. Peter 1992), they yielded a surprisingly young age of Tournaisian Zone 5 (Fig. 10; Dolby 2004).

The new date has necessitated a complete reassessment of previously held definitions of the Albert Formation. As a consequence, the grey mudstones, shales, and associated evaporites formerly assigned to the upper part of the Albert Formation — and which included the Gautreau Formation of Norman (1932), the 'Gautreau evaporites' of Greiner (1962), and the Gautreau Member of Carter and Pickerill (1985a, 1985b) — are now assigned to the Gautreau Formation of the Sussex Group. The Gautreau Formation is exposed in a tract from Anagance Ridge to The Gorge (Fig. 4, 5, 6) and was included in the Albert Formation by all earlier mappers.

The Albert Formation in the report area is exposed in the Apohaqui–Wards Creek area in the Moncton Subbasin (Fig. 4), Catamount–McQuade Brook area in the northwestern domain of the IMDZ (Fig. 6), and Lower Millstream–Cosman Settlement area in the Cocagne Subbasin (Fig. 4).

Beds of the Moncton Subbasin between Apohaqui and Wards Creek have been mapped and briefly described by St. Peter (1995a, 2002). In this area, the Albert Formation is divisible (in ascending order) into the Dawson Settlement, Frederick Brook, and Hiram Brook members, as defined by Greiner (1962). The Dawson Settlement Member comprises grey to green polymictic conglomerate, lithic sandstone, siltstone, shale, and minor organic shale. The Frederick Brook Member is dominated by organic (kerogenous) shale, siltstone, and minor sandstone (Fig. 12). The Hiram Brook Member consists of intervals up to several tens of metres thick of grey shale, kerogenous shale, and siltstone and interstratified intervals several metres or tens of metres thick of coarse to fine-grained sandstones (Fig. 13). The Hiram Brook sandstones are reservoir rocks for the natural gas found in the McCully Natural Gas Field (hereafter called the McCully Field) northeast of Sussex (Fig. 4).

The Albert Formation in the Apohaqui–Wards Creek area is generally interpreted as a lacustrine sequence (Greiner 1962). This is particularly apparent in sections within the formation that bear wave-ripple-laminated sandstones (Fig. 14) and organic-rich varved

Figure 12. Laminated kerogenous and dolomitic shale (oil shale) of the Frederick Brook Member, Albert Formation. The syn-sedimentary slump folds shown beneath the coin were commonly developed in these wet and unstable organic-rich lacustrine muds. Diameter of coin is 2.6 cm.

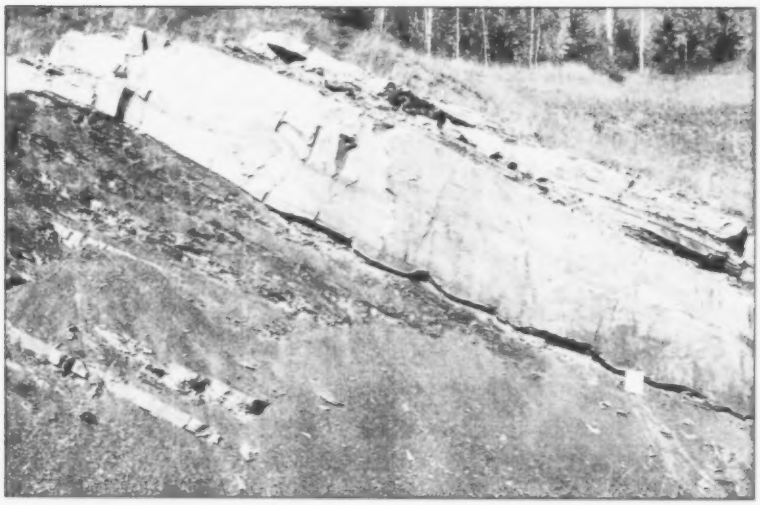

Figure 13. Interbedded, lacustrine, dark grey organic-bearing shales and light grey turbiditic (?) sandstones of the Hiram Brook Member, Albert Formation. Clipboard in lower right is 23 cm wide.

Figure 14. Wave-ripple-laminated sandstone of the Albert Formation. This feature indicates deposition in above-wave-base standing water. Hammer for scale.

shales. However, some of the fining-upward sandstone-siltstone intervals in lithologic well logs in the Hiram Brook Member from the McCully Field are several metres thick and almost certainly fluvial in origin.

In the Catamount–McQuade Brook area (Fig. 6), lower beds of the Albert Formation are generally in fault contact with the McQuade Brook Formation. As mentioned above, the Albert Formation is interpreted to concordantly overlie strata of the McQuade Brook Formation 1.5 km south of McQuade Brook, although the contact is not exposed. From Indian Mountain to Shediac River, the Albert Formation is concordantly overlain by the Indian Mountain Formation.

The Albert Formation in the Catamount–McQuade Brook area (Fig. 6) consists of brown-weathering, dark grey, calcareous and dolomitic or ankeritic, pyritic, laminated, slightly kerogenous siltstone and mudstone; minor buff-weathering, grey, fine- to coarse-grained, commonly graded, quartzofeldspathic arenite; minor grey, parallel-laminated shale; minor brown-weathering, dark grey kerogenous shale; and minor intraformational conglomerate. Intervals of sandstone, siltstone, and mudstone similar to those in the McQuade Brook Formation occur locally in the lower part of the Albert Formation. The kerogenous shales and mudstones are characterized by varved laminations of kerogen, carbonate minerals, and silicates; locally, they contain fossil fish (Greiner 1977) and fish scales. These beds and the interbedded coarser strata have been interpreted as lacustrine deposits (Greiner 1962, 1974; Smith and Gibling 1987; St. Peter 1992).

The lacustrine kerogenous shales and mudrocks so typical of the Albert Formation in the Catamount–McQuade and Apohaqui–Wards Creek areas have not been observed in the formation in the Lower Millstream–Cosman Settlement area (Cocagne Subbasin). Instead, the Albert section there comprises a series of stacked, fining-upward units of olive green to grey, polymictic granule, pebble, and rare cobble conglomerate; very coarse- to medium-grained, flat- or cross-bedded feldspathic sandstone; fine-grained to silty rippled sandstone; and mudstone or shale (Fig. 15). Conglomerate is a relatively minor part of the formation here and typically is clast supported. Most clasts are rounded and comprise competent lithotypes of granitic and felsic volcanic rocks. The finer sandstones and mudrocks commonly have abundant carbonized plant detritus and locally display rooted intervals. The sandstone in this area is characterized by abundant feldspar, and sandstone beds in the lower and upper parts of the formation are commonly pinkish grey. These features indicate that the formation has a gradational relationship with both the underlying, feldspar-rich Memramcook Formation and the overlying Bloomfield Formation.

Spore assemblages were recovered from 14 samples of the Albert Formation in the Catamount–McQuade Brook area. Most of those assemblages indicated a Tournaisian Zone 2 age (Fig. 10). Two samples yielded a somewhat more precise possible age of Zone 1D to

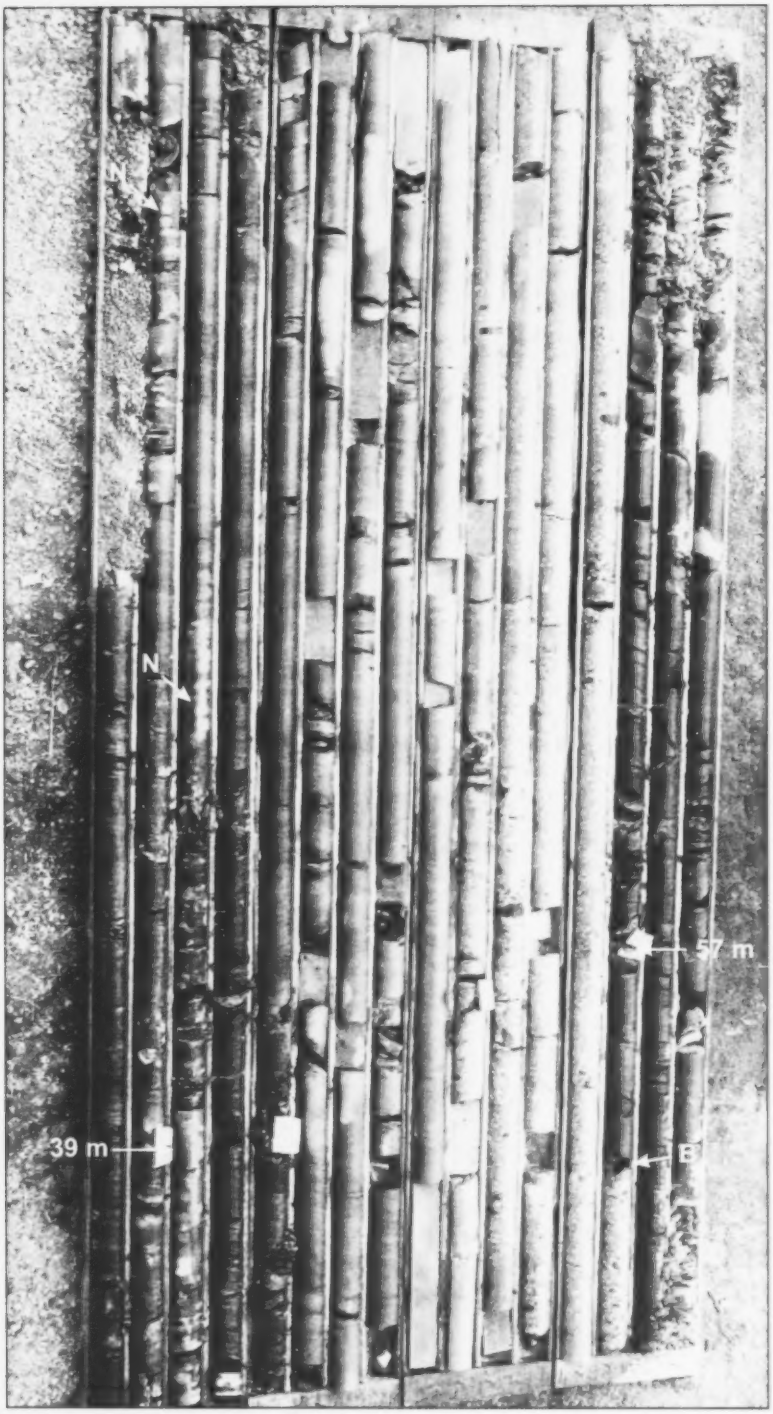

Figure 15. Core of the Albert Formation from Gulf Minerals Canada Ltd. corehole Lower Millstream No. 1. Shows typical fining-upward cycle of pinkish grey pebble conglomerate and coarse sandstone overlain by grey rippled sandstone and capped by dark grey, laminated to massive mudstone. Base of the cycle (B) is erosive on the underlying dark grey shales. Note the incipient light grey micrite nodules (N) in shales at the top of the cycle. Wooden markers indicate depth in metres below surface. Core diameter is 3.5 cm.

Zone 2A; and six samples gave a general Zone 2 or 3 age (Dolby 1991, 1995a, 1995b, 2004, 2006a). Grey mudrocks that potentially could yield spores in the Millstream–Cosman Settlement area are rare. From the twelve samples collected there, five gave a general Tournaisian age, four gave a more precise Tournaisian Zone 1D to 2 age, and three implied a somewhat younger Zone 2 to 3A age (Dolby 1995a, 2001b, 2006b).

Indian Mountain Formation

Indian Mountain Formation is a new name for a unit of clastic rocks that conformably overlies the Albert Formation in the northwestern domain of the IMDZ. The new formation is exposed in a northeasterly trending tract between the community of Indian Mountain and the Shediac River headwaters and is bounded to the south by the North River Fault (Fig. 6). Historically, the beds herein assigned to the Indian Mountain Formation were included in the Moncton Group of Norman (1941b), Gussow (1953), and Carr (1968) or in the Weldon Formation of Johnson and St. Peter (1997).

The conformable contact between the Indian Mountain and Albert formations is exposed 1.5 km east of Indian Mountain in a small quarry on the east side of a stream that drains south into North River. In this quarry and along the road leading to it, beds at the base of the Indian Mountain Formation are olive green to buff, pebbly to medium-grained sandstones with interbeds of fine-grained, rippled sandstone and minor beds of mudstone and shale (Fig. 16). The sandstone beds locally have prolific plant fragments, including *Sigillaria*. In small outcrops along the stream south of the quarry, the olive sandstones that characterize the lower part of the formation give way upsection to reddish brown, mainly fine-grained sandstones with interbedded reddish brown mudstones.

Spores from three localities in the Indian Mountain Formation yielded a nonspecific Tournaisian Zone 2 to Zone 3 age (Dolby 1995a, 2006a). The Indian Mountain Formation must be either upper Zone 2 or lower Zone 3 age, because it lies conformably above the Albert Formation, which palynologically is more or less constrained to a Zone 2 age (Fig. 10).

Bloomfield Formation

The newly named Bloomfield Formation refers to a sequence of redbeds that extends in a northeast-trending tract between Searsville and Springhill in the southwestern part of the Cocagne Subbasin (Fig. 4). The type area occurs in the Moncton Subbasin near the community of Bloomfield 25 km southwest of Sussex (Fig. 2). There, kerogenous mudstones and interstratified sandstones of the Albert Formation are gradationally overlain by grey to green and reddish brown shales and mudrocks representing the basal part of the Bloomfield Formation (Fig. 17). These fine-grained beds are succeeded by a coarsening-upward

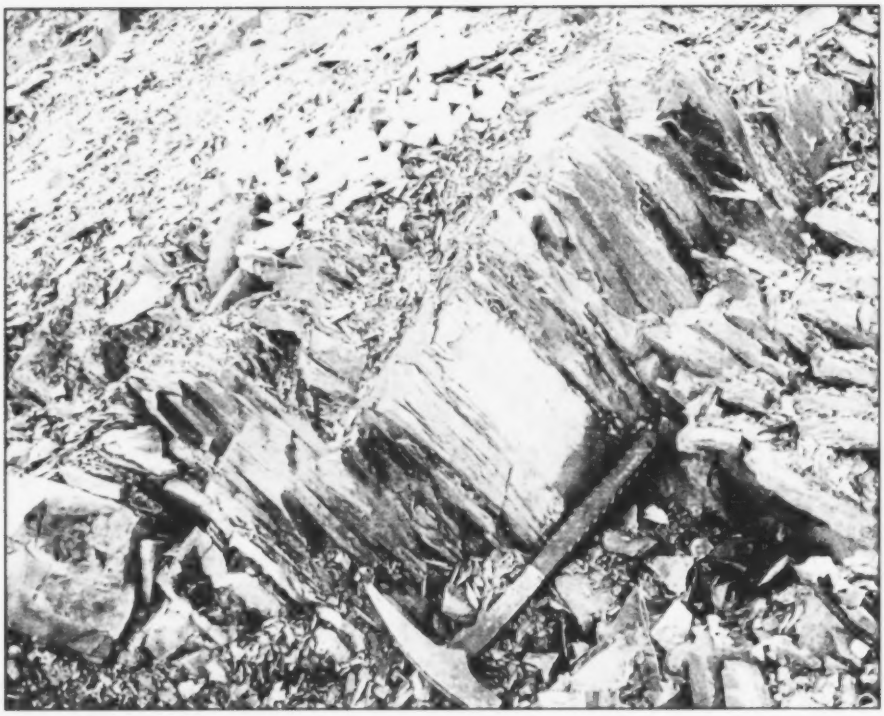

Figure 16. Strata of the Indian Mountain Formation exposed in a small quarry east of Indian Mountain (Fig. 6) showing typical olive green, fine-grained sandstone to siltstone with very thin partings of green shale.

Figure 17. Outcrop of the Bloomfield Formation in an old quarry in the type area near Bloomfield (Fig. 2). Photograph shows lower beds of the formation: reddish brown shale, and rare grey shale (near the top) with thin interbeds of buff to grey sandstone. Hammer for scale.

Figure 18. Outcrop of the Bloomfield Formation along Mill Brook northeast of Lower Millstream. Photograph shows typical friable pebble to granular conglomerate, feldspar-rich sandstone, and lenticular beds of red mudstone. Hammer for scale.

interval of pink, coarse-grained feldspathic sandstones and pebble conglomerates with lesser interbeds of cobble conglomerate and minor mudrocks (Fig. 18). This coarse interval is succeeded by an interval dominated by reddish brown mudrocks and minor sandstones (St. Peter 1989).

The Bloomfield Formation gradationally overlies the Albert Formation in the Searsville–Springhill tract, where it comprises a series of stacked, fining-upward units of pink, friable pebble conglomerate; pink to reddish brown, friable, medium-grained to granular feldspathic sandstone; and lesser reddish brown to grey or green siltstone and mudstone. The pebble conglomerate and coarse-grained sandstone contain abundant pink feldspar and undoubtedly were derived from a granitic source. Locally, the upper fine-grained parts of the cycles contain grey mudstones, organic-rich (kerogenous) mudstones, and subordinate nodular and oncolitic (?) carbonate. Although previously assigned to the Albert Formation, these organic-bearing and carbonate beds have been described in some detail by Carter and Pickerill (1985b) and ascribed an algal swamp depositional environment. This interpretation is consistent with the clearly fluvial environment of deposition implied by the stacked, fining-upward conglomerate-sandstone-mudstone cycles.

Six spore-bearing samples from the Bloomfield Formation (three from cores and outcrop in the Berwick–Lower Millstream area (Fig. 4) and three from the type area near Bloomfield) were dated by Dolby (2001b, 2005, 2006b) and indicated a Tournaisian Zone 2 or 3A age (Fig. 10).

Sussex Group

Sussex Group is a name introduced by St. Peter (2002) for a sequence of continental clastic rocks that unconformably overlies the Horton Group (Fig. 19) and unconformably or disconformably underlies the Windsor Group in the Sussex area of the Moncton Subbasin. Prior to St. Peter's surface mapping, the relationship of the Sussex Group to its bounding rocks was recognized by Corridor Resources Inc. from seismic profiles and exploration boreholes in the McCully Field (Fig. 4).

Figure 19. Unconformity (dashed line) between dark grey shales of the Albert Formation (Horton Group) and overlying conglomerates of the Mill Brook Formation (Sussex Group). Note the light grey clay mineral(s) being washed out and down the rock face, coating the shales immediately below the base of the unconformity. Outcrop lies along Highway 1 between Sussex and Apohaqui.

In the type area of the Sussex Group in the Moncton Subbasin, the group is divided into the lower Mill Brook and upper Dutch Valley formations. In the Cocagne Subbasin, the group consists of a single unit: the newly named Ridge Brook Formation (Fig. 4, 5). In the southeast domain of the IMDZ (Fig. 6, 7), the group comprises the Stilesville Formation (new name), Gautreau Formation (redefined), Weldon Formation, and Briggs Cross Formation (new name).

Mill Brook and Dutch Valley Formations

In the type area, the Sussex Group comprises coarse conglomerates, lithic sandstones, and minor mudrocks of the Mill Brook Formation; and gradationally overlying, mainly red mudstones with subordinate sandstones and rare conglomerates of the Dutch Valley Formation. These formations are described in some detail by Hinds and St. Peter (2006, this volume) and thus are not discussed further in the present paper.

Ridge Brook Formation

The newly named Ridge Brook Formation designates a coarse-grained unit of the Sussex Group that is widely distributed in the Cocagne Subbasin (Fig. 4, 5, 6). The formation is exposed on the northwest side of the Smith Creek Fault from Searsville in the southwest to Catamount in the northeast.

The stratigraphic position of the rocks herein assigned to the Ridge Brook Formation has long been debated, in part because they consist nearly exclusively of coarse-grained redbeds that (until work undertaken by the present author) have yielded no fossils, and in part because their contact relationships with underlying and overlying rocks have not been adequately studied or understood. In the past, these Ridge Brook strata were either regionally or locally placed in the Moncton Group (Stewart 1941a, 1941b, 1941c; Alcock and MacKenzie 1946; Gussow 1953; Hamilton 1961), the Hillsborough Formation of the Moncton Group (Gussow 1953; Hamilton 1962a), or a coarse facies of the Moncton Formation (St. Peter 1989).

The Ridge Brook Formation is juxtaposed against basement along the Belleisle Fault at Ridge Brook north of Havelock. Adjacent to the basement, beds of the formation dip gently southeast but otherwise show little evidence of disruption by the fault. However, the absence of the formation on the basement inlier at Salem attests to the substantial magnitude of displacement between rocks of the basement and Ridge Brook Formation along the fault. Carbonates at the base of the Windsor Group overlie the formation in both the Havelock and the Lower Millstream areas. The Ridge Brook Formation is older than the unconformably overlying rocks of the Windsor Group, including the Hillsborough Formation. On the basis of these relationships, the Ridge Brook Formation is assigned to the Sussex Group.

The upper contact of the Ridge Brook Formation with the base of the Gays River Formation (Windsor Group) is interpreted as a disconformity. Little evidence exists for an angular discordance between the Ridge Brook and Gays River formations: in the Havelock area, for example, bedding in both formations is gently dipping to horizontal. Nonetheless, the fact that the Gays River Formation overlies the Ridge Brook Formation at Havelock, yet overlies basement rock east of Salem, suggests an overstepping unconformity. The contrast in ages of the Ridge Brook (Tournaisian) and Gays River (Visean) formations also supports the inference of a hiatus (see below).

The type section of the Ridge Brook Formation along Ridge Brook is about 600 m thick and includes exposures that reach from the basement rocks west of Harewood to the first Windsor limestone bed north of Havelock (Fig. 4). Outcrops there occur fairly continuously in the stream channel and along outside turns of Ridge Brook, and are best observed at low summer water. Neither the basal nor upper contact of the formation is exposed at the type section. It is not known how much additional section lies in the subsurface beneath exposed beds on the southeast side of the Belleisle Fault. The formation shows a subtle overall fining upward from dominantly pebble and cobble conglomerate in its lower part to interstratified conglomerates, pebbly grits, sandstones, and mudstones in its upper part.

Concomitant with the upward fining, rocks in the lower and upper parts of the formation exhibit somewhat contrasting colours, textures, and mineralogies. Conglomerates in the lower part typically are drab maroon and have a muddy sand matrix. These conglomerate and sandstone beds contain abundant, typically angular clasts of grey slate, siltstone, and sandstone that are analogous to Silurian metasedimentary rocks of the Kingsclear Group (New Brunswick Department of Natural Resources and Energy (NBDNRE) 2000) exposed just to the north on the New Brunswick Platform (Fig. 2, 4).

In contrast, the conglomerates, pebbly grits, and sandstones in the upper part of the formation are more commonly pink and friable. They show lenticular bedding and, locally, cross-bedding. Associated minor beds of fine-grained sandstone and mudstone are red to reddish brown and occur as parallel beds that commonly are partially or completely cut out by overlying conglomerate or sandstone intervals. These conglomerate and sandstone beds contain greater quantities of granite, feldspar, and quartz. They also include subordinate clasts of quartz-feldspar porphyry, felsic volcanic, mafic volcanic, and mafic plutonic rocks. Some fine-grained strata contain micritic calcite nodules.

Cross-beds in the Ridge Brook type section and elsewhere in the Havelock–Killams Mills area show a generally southeasterly transport direction for the Ridge Brook detritus. This evidence, together with the clast composition, strongly indicates that sediment was sourced from metasedimentary rocks of the Kingsclear Group and granitic plutons that lie north and northwest of the report area (NBDNRE 2000). A northern sediment source implies that the Belleisle Fault system approximately represents the northwestern depositional margin of the Sussex Group in the Cocagne Subbasin (Fig. 2).

On the south side of the basement block at Jordan Mountain, the Ridge Brook Formation dips southeast and shows a general fining upward. It has abundant conglomerates that contain clasts up to small boulder size in the lower beds, but it features more pebbly grits, sandstones, and minor mudstones in the upper beds (Fig. 20). The clasts are dominated by granite, quartz-feldspar porphyry, feldspar porphyry, and gabbro, all common rock types in the adjacent basement. In outcrops adjacent to the Jordan Mountain basement, massive beds of conglomerate contain angular, basement-derived clasts in a sandy mud matrix. These beds are almost certainly debris flows.

East of Mount Middleton and just east of the Jordan Mountain Fault, polymictic conglomerates of the Ridge Brook Formation carry a few rounded pebbles of olive green, medium- to coarse-grained sandstone that most likely were eroded from rocks of the Albert Formation situated immediately west of the fault (Fig. 4). The pebbles imply the existence of an erosional unconformity between the Ridge Brook Formation and underlying Horton Group.

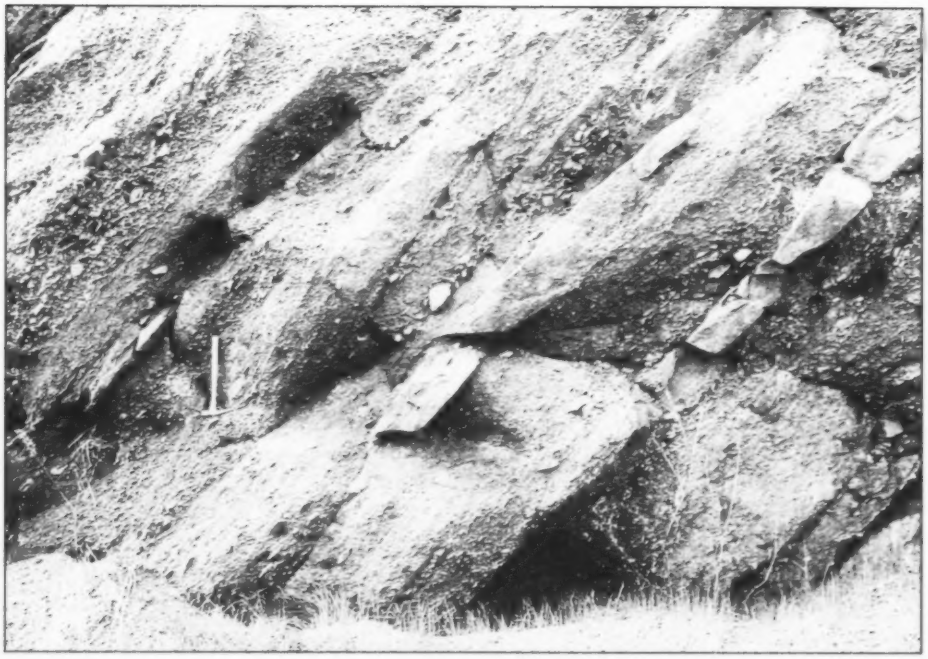

Figure 20. Outcrop of the Ridge Brook Formation in a highway cut southwest of McGregor Brook (Fig. 5). Photograph shows cross-bedded, polymictic cobble-pebble conglomerates and lenticular interbeds of sandstone. Hammer for scale.

Between Killams Mills and Monteagle (Fig. 5), several exposures of conglomerates and sandstones of the Ridge Brook Formation are greenish grey to nearly black. The colour results from pervasive replacement of the matrix by manganese oxides. The black, manganiferous beds previously were misidentified as dead oil or bitumen 'shows' (St. Peter 1995b) and, in 1941, even led to the drilling of an oil and gas exploration hole by New Brunswick Gas and Oilfields Ltd. at Killams Mills.

The near-pervasive oxidized state (red colour) of the Ridge Brook Formation precludes the preservation of palynomorphs. However, one assemblage was recovered from an interval of grey to black shale intersected at a depth of 110 m to 121 m below surface in the Columbia borehole L-18 near Monteagle (Fig. 5). Although the assemblage was sparse, it contained *Crassispora trychera* and supported a Tournaisian age of no older than Zone 3B (Fig. 10; Dolby 2001a).

Stilesville Formation

The newly named Stilesville Formation refers to a unit of coarse-grained clastic rocks that defines the base of the Sussex Group in the southeastern domain of the IMDZ in The Gorge area (Fig. 6). Because the unit is areally restricted, the map scale of Figure 6 does not allow for a complete delineation of the formation. Stilesville beds, for example, lie just north and south of the basement block immediately north of The Gorge. Also, the triangular wedge of Stilesville Formation just east of The Gorge contains an exhumed core of basement rock in the eastern part of the wedge. Readers seeking more detailed maps of this formation may consult St. Peter and Park (2006a, 2006b, 2006c).

At its type area near The Gorge, rocks of the Stilesville Formation are well exposed in several stone quarries dug mainly in basement rocks, where it lies unconformably and in fault contact with the highly faulted basement. The Stilesville Formation comprises reddish brown and greyish green, angular to subrounded granule to boulder polymictic breccia and conglomerate with varied quantities of clasts of granite, diorite, schist, feldspar, and recycled sandstone from the Horton Group (Fig. 21, 22). Greyish green breccia and megabreccia (regolith) with angular clasts and blocks of basement lithotypes up to several metres long occur adjacent to the basement exposures. Massive breccia beds are commonly difficult to distinguish from the cataclastic basement granitoids from which they are derived. Minor components of the formation consist of reddish brown and greyish green, fine-grained to granular, lenticular-, parallel-, and rarely cross-bedded sandstone; minor reddish brown and grey siltstone and mudstone; and very rare silcrete.

Beds of the Stilesville Formation were deposited on a basement that apparently had considerable relief, as suggested by the formation's wide range in thickness from near-zero to 250 m. Their coarse grain size and massive nature suggest that the Stilesville beds are primarily debris flows, and that the formation represents a proximal alluvial-fan sequence.

Four dateable palynomorph assemblages were recovered from grey mudrocks of the Stilesville Formation and yielded a late Tournaisian Zone 3B to 5 age (Fig. 10; Dolby 1995b, 2006a).

Gautreau Formation

Norman (1932) introduced the name Gautreau Formation to define a unit of grey mudrocks, sulphates, and rock salt that occurs in the Moncton Subbasin near Hillsborough (Fig. 2). His original stratigraphic definition was refined by Greiner (1962), St. Peter and Johnson (1997), and Park and St. Peter (2005). Park and St. Peter (2005) also provided evidence for assigning the Gautreau Formation to the Sussex Group.

In the present report area, beds of the Gautreau Formation are exposed in the southeastern domain of the IMDZ, forming a continuous, 45 km tract from Anagance Ridge in the southwest to The Gorge in the northeast (Fig. 4, 5, 6). This tract of dominantly grey mudrocks previously was included in the Albert Formation (Stewart 1941a, 1941b, 1941d; Norman 1941b; Gussow 1953; Greiner 1962; Carr 1968; St. Peter 1989).

Total thickness of the Gautreau Formation cannot be determined. The formation's lower contact is faulted from Anagance Ridge (Fig. 4) to The Gorge (Fig. 5, 6). At The Gorge, the basal boundary is in part undisturbed, but higher beds in the unit are faulted out. The formation is interpreted to thin toward basement rocks exposed at The Gorge, where it generally lies gradationally above the Stilesville Formation. Where the Stilesville Formation

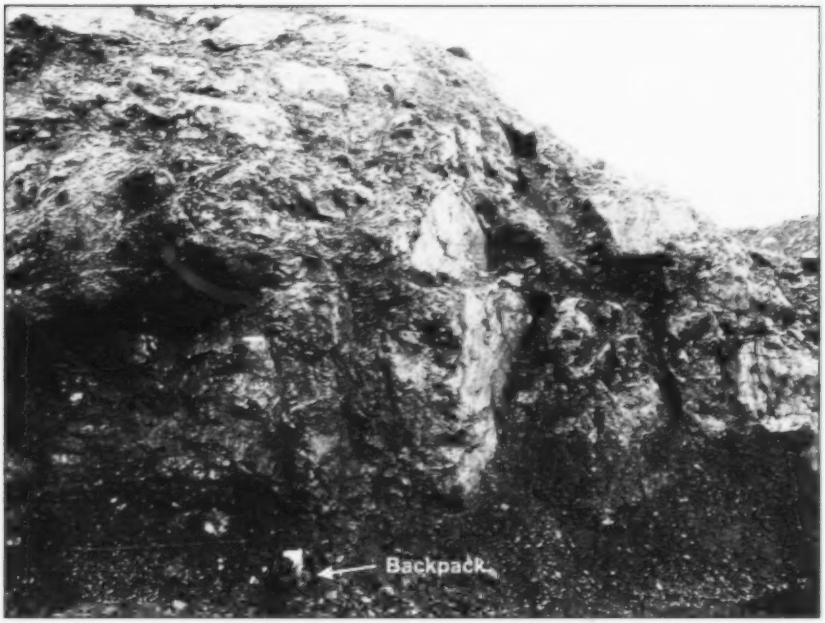

Figure 21. Outcrop of the Stilesville Formation in the Dexter Quarry at The Gorge (Fig. 6). Photograph shows highly fractured megabreccia comprising abundant angular blocks of locally derived pink granite set in a dark grey heterogeneous matrix of mud to cobbles. Backpack and clipboard for scale.

Figure 22. Outcrop of the Stilesville Formation in the Moncton Crushed Stone Quarry at The Gorge (Fig. 6). Photograph shows typical conglomerate with locally derived angular clasts of pink cataclastic and mylonitic granite, darker grey diorite, white vein quartz, and subtle, recycled reddish brown sandstone (see arrows) from the Horton Group.

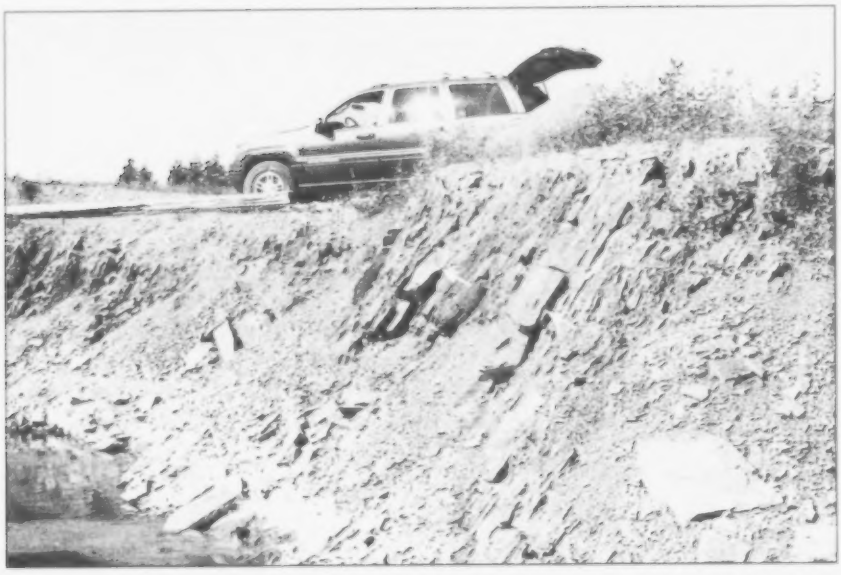

Figure 23. Strata of the Gautreau Formation, northwest of Lutes Mountain (Fig. 5, 6), showing interbedded grey and reddish brown shales with thin interbeds of rusty brown sandstone (arrows).

is locally absent, however, the Gautreau Formation lies unconformably on basement granitoid rocks (St. Peter and Park 2006a). The Gautreau section south of Monteagle along North River is largely concealed but may be up to 1500 m thick.

In The Gorge area, the Gautreau Formation is gradationally overlain by the Briggs Cross Formation. Farther southwest, along North River and as far south as Anagance Ridge, the Gautreau Formation is conformably succeeded by fine-grained redbeds of the Weldon Formation. The Gautreau–Weldon formational boundary is placed where the dominant colour changes from grey (Gautreau Formation) to reddish brown (Weldon Formation).

In the IMDZ, the Gautreau Formation comprises mostly grey shale and mudstone or silty mudstone with lesser thin beds of fine- to medium-grained sandstone. In the northeastern part of the zone, the formation contains significant reddish brown shale and mudstone (Fig. 23) and also interbeds of coarse-grained sandstone and rare polymictic conglomerate. Both the grey and reddish brown mudrocks typically are somewhat calcareous and locally host calcrete nodules and rare nodular limestone beds. In 1973 a borehole sunk by the DNR (Province of New Brunswick, Cornhill No. 3) in the Cornhill area (Fig. 4) to a depth of 944 m penetrated a thick section of Gautreau mudrocks. The section included beds of rock salt and rock salt/mudstone breccia (Crosby 1975).

Carter (1985) and Carter and Pickerill (1985b, described beds of the Gautreau Formation in the Moncton Subbasin and IMDZ and, at the time of their writing, considered the strata as part of the Albert Formation. The authors reasoned that the beds were deposited on a broad mudflat containing restricted evaporitic lakes. St. Peter (1993) cited evidence that Gautreau halite shows textural features of primary lacustrine precipitate and also later diagenetic and epigenetic mineral growth.

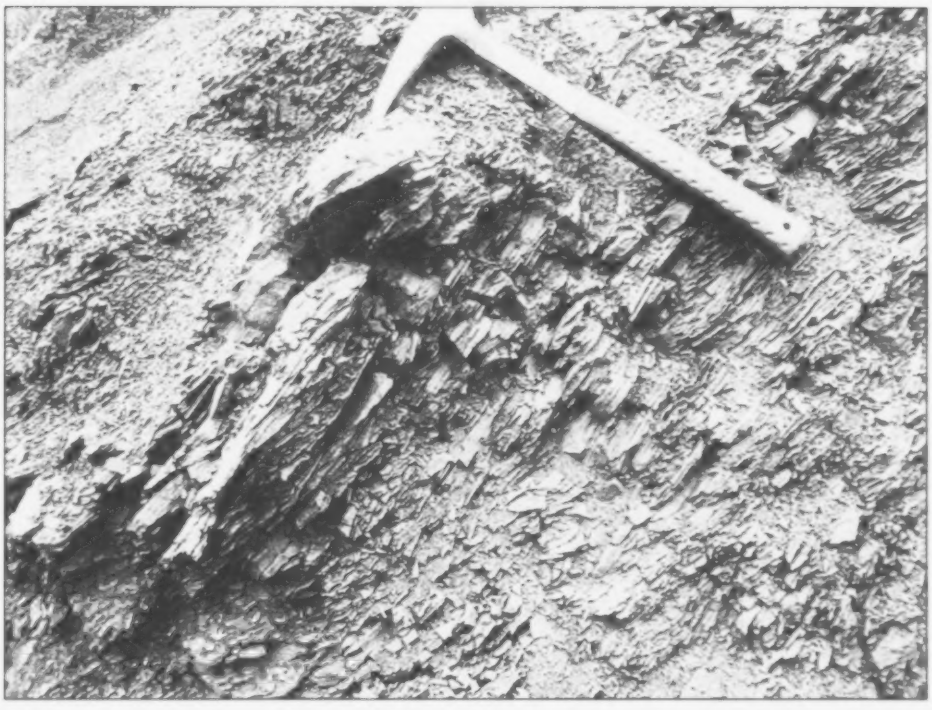

Figure 24. Exposure of the Weldon Formation along Salt Springs Brook near Intervale (Fig. 5), showing typical reddish brown shale with thin interbed of rippled sandstone.

Spore assemblages from 19 localities in the Gautreau Formation from Anagance Ridge to The Gorge yielded a consistent late Tournaisian Zone 3B to 5 age (Fig. 10; Dolby 1995a, 1995b, 2006a). The late Tournaisian age of the formation contrasts with the early Tournaisian age of the Albert Formation. This age difference and the strong lithological contrasts between the two formations together provide solid justification for separating the Gautreau Formation from the Horton Group and placing it in the Sussex Group.

Weldon Formation

The Weldon Formation in the IMDZ gradationally overlies the Gautreau Formation in a northeast-trending, wedge-shaped body that extends for 35 km from the headwaters of North River through Intervale and Scott Road to Anagance Ridge (Fig. 4, 5). From near the headwaters of North River to Intervale, the Weldon Formation shows a remarkable southwestward thickening, passing from near-zero at its northeastern limit to greater than 2 km at Intervale. The thickness estimate at Intervale assumes that the section, which is intermittently exposed along North River, is not repeated by unknown faults. The Weldon Formation is also exposed in a second northeast-trending tract that reaches from Smith Creek to beyond Cornhill (Fig. 4, 5).

The Weldon Formation comprises reddish brown mudstone and shale; and minor reddish brown to grey, parallel- and cross-laminated, fine-grained sandstone (Fig. 24). Interbeds of grey mudstone and shale, coarse-grained sandstone, polymictic conglomerate, and

limestone are relatively rare. The mudstones and shales commonly show mudcracks and rarely rainprints. Carbonate nodules (calcrete) are also present in some mudstone intervals. The preponderance of desiccated mudstones and shales with associated calcrete suggests that the Weldon Formation was deposited on a fairly dry but extensive and persistent alluvial plain, and that it represents either fluvial-overbank and/or distal alluvial-fan deposits (Reading 1978; Miall 1992).

Nine dateable spore assemblages were recovered from the Weldon Formation, and, as with assemblages from the Gautreau Formation, they indicated a late Tournaisian Zone 3B to 5 age (Fig. 10; Dolby 1995a, 1995b, 1996a).

On the southeast side of the Smith Creek Fault, the Weldon Formation underlies carbonates at the base of the Windsor Group near Smith Creek and Intervale. On the northwest side of the same fault, however, the Ridge Brook Formation underlies basal Windsor carbonates at Harewood, Havelock, and near Lower Millstream and Parleeville (Fig. 4, 5). Given these contact relationships with the Windsor Group, and given the fairly equivalent ages (Tournaisian Zone 3B to 5) of the Ridge Brook and Weldon formations, the two formations can be interpreted to represent coeval proximal and distal alluvial depositional environments, respectively, within the Sussex Group. In this context, it can be inferred that juxtaposition of the proximal Ridge Brook Formation and the distal Weldon Formation resulted from considerable displacement along the Smith Creek Fault.

In the Sussex area of the Moncton Subbasin (Fig. 4), fine-grained redbeds that immediately underlie the Windsor Group and occupy a stratigraphic position analogous to the Weldon Formation are called the Dutch Valley Formation (St. Peter 2002). The Dutch Valley Formation is described in Hinds and St. Peter (2006, this volume).

Briggs Cross Formation

Briggs Cross Formation is a new name assigned to the predominantly coarse-grained redbeds that lie concordantly above the Gautreau Formation in The Gorge area (Fig. 6) and above the Weldon Formation from near the headwaters of North River to Scott Road (Fig. 5, 6). The type area of the Briggs Cross Formation lies about 1 km southwest of The Gorge along a side road on the south side of Briggs Cross Road (St. Peter and Park 2006a, 2006c). Another good reference section occurs along North River about 7 km southwest of Catamount and exposes the gradational contact with the underlying Weldon Formation.

Historically, the strata herein assigned to the Briggs Cross Formation have been assigned in whole or in part either to the Moncton Group (Stewart 1941a; Norman 1941b), or to an undesignated boulder conglomerate unit of the Horton Group or Moncton Group (Carr 1968), or to a coarse facies of the Weldon Formation (St. Peter 1995b–1995g). Gussow

(1953) appears to have been perplexed by this unit, because he placed different parts of it in the Horton Group (Memramcook Formation), Cumberland Group (Boss Point Formation), and Pictou Group (Salisbury Formation). The Briggs Cross Formation consists nearly exclusively of coarse-grained redbeds and consequently has rarely been sampled for palynomorphs. This lack, coupled with inadequate study of the contact relationships with underlying and overlying rocks, has led to the conflicts in interpretation.

The Briggs Cross Formation has a gradational boundary with the underlying Weldon Formation. It thereby is distinguished from the conglomerate-dominated Hillsborough Formation (base of the Windsor Group), which lies with angular unconformity on the Weldon Formation in the Moncton Subbasin (Wright 1922; Park and St. Peter 2005). Consequently, the Briggs Cross Formation almost certainly constitutes a unit assignable to the Sussex Group and is older than the unconformably overlying Windsor Group, including its basal Hillsborough Formation.

The Briggs Cross Formation comprises mostly reddish brown (and, rarely, greyish green), polymictic granule to cobble conglomerate with assorted clasts of subangular to rounded granite, diorite, schist, feldspar, and recycled Carboniferous sandstone (Fig. 25, 26, 27). The conglomerates contain interbeds of reddish brown and grey to green, fine-grained to granular, lenticular, parallel-bedded and rarely cross-bedded sandstone; and minor reddish brown and grey siltstone and mudstone. Larger exposures typically show either multiple stacked sequences of conglomerate or beds of conglomerate, sandstone, and mudstone arranged into fining-up cycles a few metres thick.

The clasts of granite, deformed granite, granitoids, and feldspar in the Briggs Cross conglomerates contrast markedly with the clasts of metasedimentary, volcanic, and granitic rocks in conglomerates of the Ridge Brook Formation. The Briggs Cross detritus is dominated by granite and deformed granite and is interpreted to have been shed from a basement complex such as that exposed at The Gorge. The wedge-shaped geometry of the Briggs Cross Formation between The Gorge and Intervale also implies a provenance associated with basement rocks exposed in The Gorge area. This interpretation is supported by the fact that both the clast size and the volume of conglomerate in the formation decrease from northeast to southwest.

The bedforms, traction structures, and dominantly coarse-grained lithotypes strongly suggest that the Briggs Cross Formation represents a proximal alluvial-fan complex comprising mainly waterlain, braided-stream deposits. The formation is probably greater than 1 km thick near The Gorge; however, it is not possible to accurately estimate a thickness at this location. Although the base of the formation is exposed along North River southwest of Catamount, the upper part of the section is faulted out by the Salt Springs Brook Fault (Fig. 5). Farther southwest, at Intervale, the Briggs Cross Formation thins out

Figure 25. Outcrop of the Briggs Cross Formation in an abandoned quarry at The Gorge (Fig. 6). Photograph shows an overturned section (tops to the left) of reddish brown, polymictic pebble to small cobble conglomerate overlain by interbedded reddish brown and grey mudstone and fine-grained sandstone. Hammer for scale.

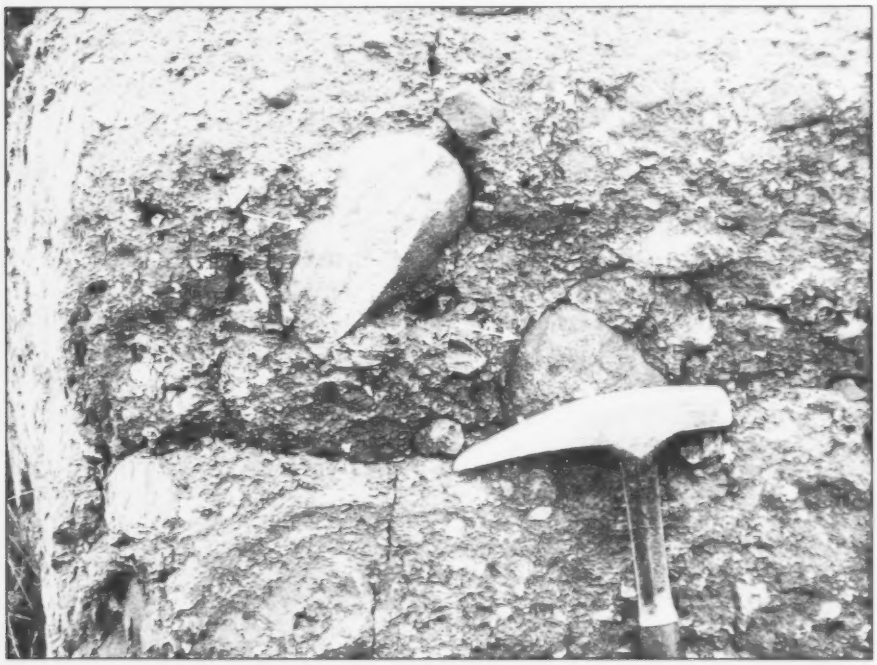

Figure 26. Outcrop of the Briggs Cross Formation, showing conglomerate with clasts of pink granite, grey diorite, and rounded cobbles (arrows) of recycled Horton sandstone.

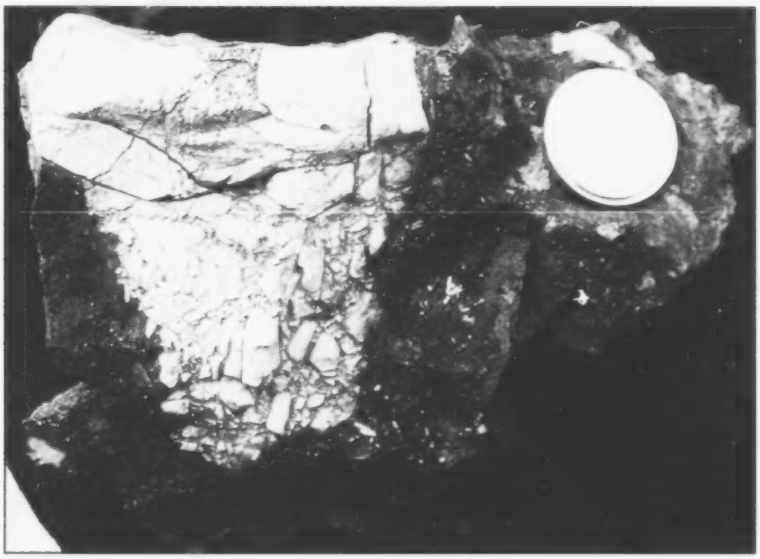

Figure 27. Sample of the Briggs Cross Formation, showing conglomerate with recycled clast of light grey, brecciated mudstone interpreted to be sourced from the Albert Formation. The relative incompetence and brecciation of the clast attests to its very local derivation. Diameter of coin is 2.3 cm.

entirely, as evidenced by the fact that the Weldon Formation is overlain by basal carbonates of the Windsor Group.

Eleven palynomorph-bearing samples were recovered from the Briggs Cross Formation. Eight samples yielded ages ranging from Tournaisian Zone 3B to Zone 5, and three samples gave a probable age of Tournaisian Zone 1 (Fig. 10; Dolby 1991, 1995b, 1996a, 1996b, 2001a, 2006a). The three older ages are believed to be the result of spores from the older, eroded strata of the Horton Group being recycled into the younger Briggs Cross beds. The common occurrence of Horton sandstone clasts in Briggs Cross conglomerates is corroborative evidence of recycling and of the existence of an unconformity between the Horton and Sussex groups.

Windsor Group

The Windsor Group in the Moncton and Cocagne subbasins and the IMDZ comprises seven formations: basal clastic rocks of the Hillsborough Formation; overlying carbonates of the laterally equivalent Gays River, Macumber, and Parleeville formations; succeeding sulphates of the Upperton Formation; chlorides of the Cassidy Lake Formation; and capping chlorides, sulphates, and carbonates of the Clover Hill Formation (Fig. 3).

The Windsor Group has not been examined in detail by the author and is given only cursory review here. McCutcheon (1981) provides the most complete account of formations within the group in New Brunswick, describing their type sections, distribution, lithotypes, environments of deposition, and ages. Additional details are given in Anderle et al. (1979);

Kingston and Dickie (1979); Roulston and Waugh (1981); Wilson (2005); Wilson et al. (2006); and Hinds and St. Peter (2006, this volume).

The carbonate formations at the base of the Windsor Group represent the initial marine transgression that extended over most of the Maritimes Basin from Newfoundland to New Brunswick (Fig. 1). These basal carbonates and younger units of the Windsor Group occur in the Cocagne Subbasin, IMDZ, and Moncton Subbasin. The early part of the subsidence cycle that ultimately gave rise to the marine invasion is reflected by the coarse, terrestrial, clastic redbeds of the Hillsborough Formation, which is found only in the latter subbasin (Fig. 4, 5).

The map legends for Figures 4 and 5 do not divide the Windsor Group, since thicknesses of the individual formations are insufficient to be shown separately at the given map scale. Instead, general information on the formation distributions is provided below.

Hillsborough Formation

The type area of the Hillsborough Formation occurs in the Moncton Subbasin near the village of Hillsborough, about 20 km south of Moncton (Fig. 2). The coarse Hillsborough redbeds there have been described by Wright (1922), Norman (1941a), and Gussow (1953). The formation outcrops east and southeast of Sussex in the present report area (Fig. 4), lying unconformably on the Dutch Valley Formation of the Sussex Group (Fig. 28).

The correct stratigraphic placement of the Hillsborough Formation has long been debated. Wright (1922) included the formation (his Zone No. 1) in his Hillsborough Series along with conformably overlying marine carbonates, sulphates, and interlayered terrestrial red clastics. Norman (1941a) removed the mainly marine strata from Wright's Hillsborough Series, called the basal coarse clastic rocks the Hillsborough Formation, and included them along with the unconformably underlying fine clastics of the Weldon Formation in the Moncton Group. Norman (1941a) also assigned the name Windsor Series to the marine and interstratified terrestrial tongues that lie concordantly above the Hillsborough Formation.

Gussow (1953) followed Norman and placed the Hillsborough Formation in the Moncton Group. However, he was aware that an unconformity existed with the underlying Weldon Formation and showed his ambivalence with the placement by stating: "it [the Hillsborough Formation] is definitely recognized as being post-Horton deformation in age and as such is more closely related to the Windsor...." (Gussow 1953, p. 1738).

Most subsequent workers have followed Gussow, ignoring stratigraphic and contact relationships with the overlying marine beds, and excluding the Hillsborough Formation from the Windsor Group, while at the same time not excluding stratigraphically higher terrestrial

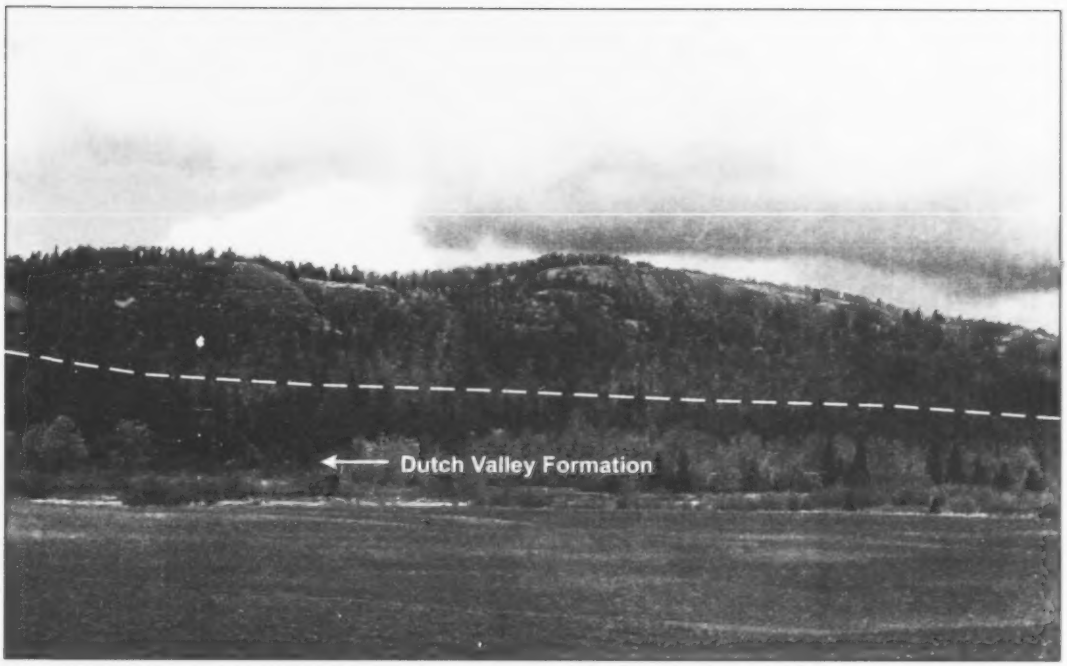

Figure 28. The Hillsborough Formation forms a prominent ridge east of Sussex at Dutch Valley, where massive conglomerates lie unconformably above recessive reddish brown mudstones of the Dutch Valley Formation. Plowed field in the foreground is founded on beds of the Dutch Valley Formation, which are visible at the toe of the cliff as 'red outcrop.' Dashed line shows approximate location of the unconformity.

tongues from the same group. This approach seems illogical, and consequently the Hillsborough Formation is considered herein to belong to the Windsor Group, as Wright (1922) originally interpreted. The base of the Windsor Group is thus defined as the unconformity at the base of the Hillsborough Formation; in the absence of the Hillsborough clastics, the base is defined as the base of the stratigraphically lowest marine carbonate formation.

Wright (1922, p. 18) described the unconformable, overstepping relationship of the Hillsborough Formation (his Zone No. 1) as follows: "red and greenish conglomerate and red arkosic sandstone; calcareous conglomerate, poorly sorted pebbles subangular, up to 2 feet in diameter, made up of pre-Carboniferous with some pebbles of Carboniferous sediments. Along Caledonia Mountain this zone thins out and frequently the overlying limestone rests on pre-Carboniferous rocks. Zone becomes thicker away from the old shoreline." Wright's assessment has not been challenged by subsequent work, and his stratigraphic assignment of the Hillsborough Formation is therefore retained here (see St. Peter 1993; and Park and St. Peter 2005).

Defining the Hillsborough Formation as the coarse clastic beds that lie concordantly below marine strata of the Windsor Group and unconformably above redbeds of the Sussex Group (or older rocks) necessarily implies restricting its distribution to the southern part of the Moncton Subbasin. For example, beds of the Ridge Brook Formation in the Cocagne Subbasin (strata that were previously placed in the Hillsborough Formation; see *Ridge*

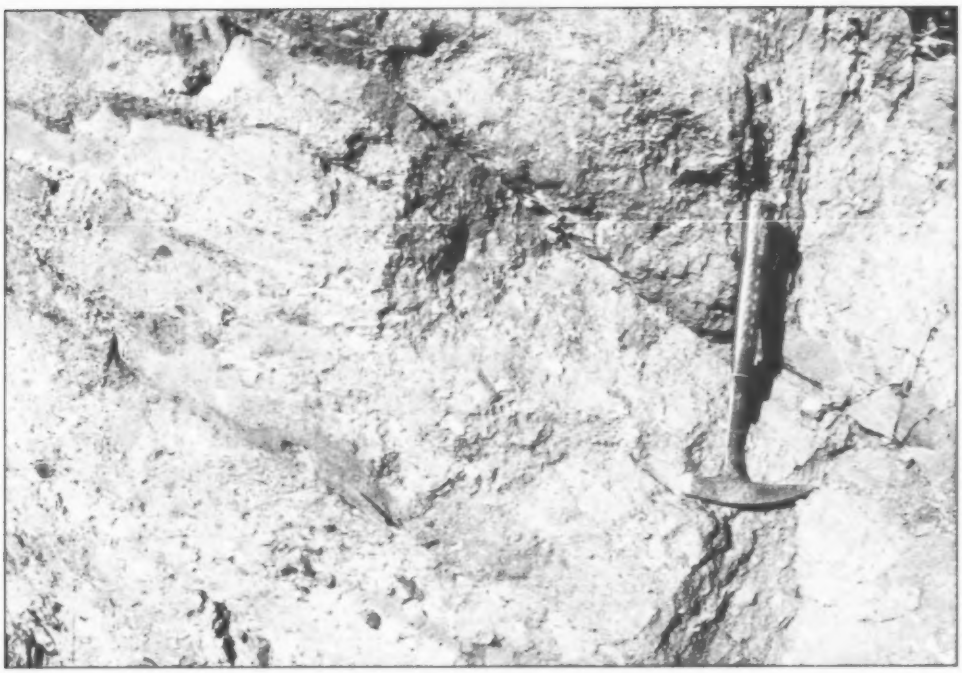

Figure 29. Beds of the Hillsborough Formation near Sussex, showing polymictic conglomerate with lenticular thin beds of reddish brown sandstone.

Brook Formation) are likely older than those of the Hillsborough Formation in the Moncton Subbasin and are disconformably overlain by basal carbonates of the Windsor Group.

The Hillsborough Formation appears to mark the axis of the subsidence that heralded the Windsor marine transgression, which would explain the wide range in thickness for Windsor marine strata. The entire Windsor marine section, for instance, is nonexistent or very thin (a few metres or tens of metres) over the Westmorland Uplift, the IMDZ, and the northern part of the Cocagne Subbasin (Fig. 2). Yet the same marine section reaches hundreds of metres thick and contains higher order evaporites in the southwestern Moncton Subbasin (McCutcheon 1981; Roulston and Waugh 1981, 1985; Wilson et al. 2006).

The Hillsborough Formation consists of reddish brown to locally grey or olive green, polymictic pebble to boulder conglomerate, conglomeratic sandstone, and minor sandstone (Fig. 29). It forms a wedge-shaped body that is 300 m thick in the south and thins out entirely over the McCully Field (Fig. 4). Hinds and St. Peter (2006, this volume) describe its lithology, contact relationships, and depositional environment. As noted by those authors, the formation's wedge shape, its coarse mass-flow and waterlain deposits, and its abundant clasts of lithotypes comparable to rocks of the Caledonia Uplift, all imply that the formation represents an alluvial-fan complex with a southern provenance.

No grey shales or mudrocks have been observed in the Hillsborough Formation, and thus no spore assemblages have been collected. Its age is interpreted as Early Visean on the basis of its stratigraphic position and the presence of the hiatus above the latest Tournaisian Ridge Brook and Weldon formations (Fig. 4, 5).

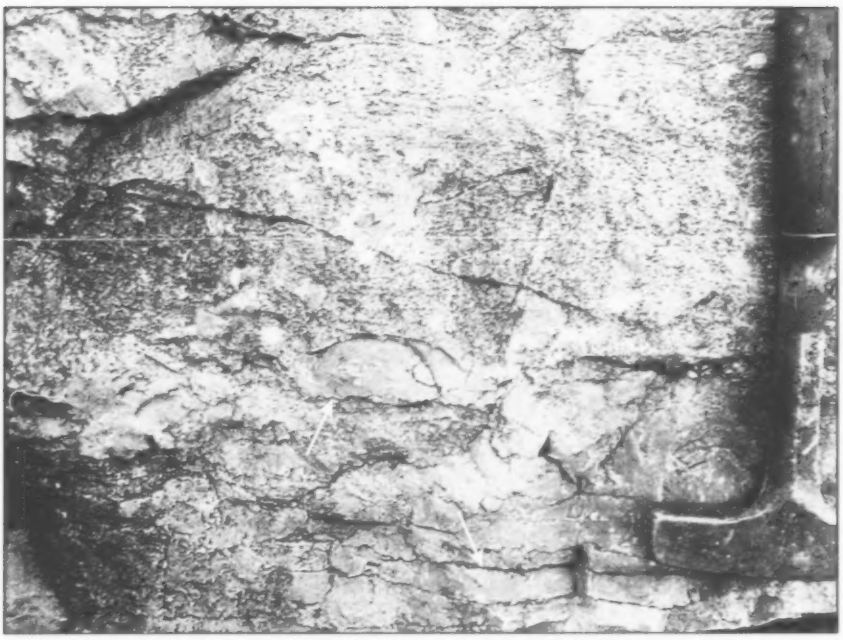

Figure 30. Beds of the Gays River Formation at a limestone quarry near Havelock (Fig. 4), showing massive boundstone with dark grey or black stylolites (arrows) that are more or less parallel to bedding.

Gays River Formation

In the Cocagne Subbasin, the Gays River Formation is a few metres to tens of metres thick and is restricted to the northern area around Havelock, Springhill, and Harewood (Fig. 4, 5). Thick- to massive-bedded algal boundstone (Fig. 30) with minor thin-bedded wackestone and packstone are the dominant rock types in this formation. The boundstones are interpreted as algal mounds (reefs) deposited in shallow water along the Windsor strand (McCutcheon 1981).

The Gays River Formation is conformably overlain by the Upperton Formation. A spore assemblage was collected from the base of the Gays River Formation in a quarry near Havelock by S. McCutcheon and examined by Barss (1979). It yielded an NS miospore zone age (Utting 1987) of late early or middle Visean (Fig. 31).

Macumber Formation

The Macumber Formation defines the basal carbonate rocks of the Windsor Group at Intervale in the IMDZ (Fig. 5), and at Smith Creek and Sussex in the Moncton Subbasin (Fig. 4). However, at Hillsborough (Fig. 2) and elsewhere in the Moncton Subbasin, the Macumber limestone lies concordantly on the Hillsborough Formation (Fig. 32), supporting the argument for retaining the Hillsborough Formation in the Windsor Group (Wright 1922). Where seen in outcrop in New Brunswick, the Macumber Formation is between 9 m and 18 m thick (McCutcheon 1981). However, it reaches about 30 m thick in natural gas exploration boreholes of Corridor Resources Inc. and Potash Corporation of Saskatchewan (PCS) Inc. in the McCully Field area (Fig. 4).

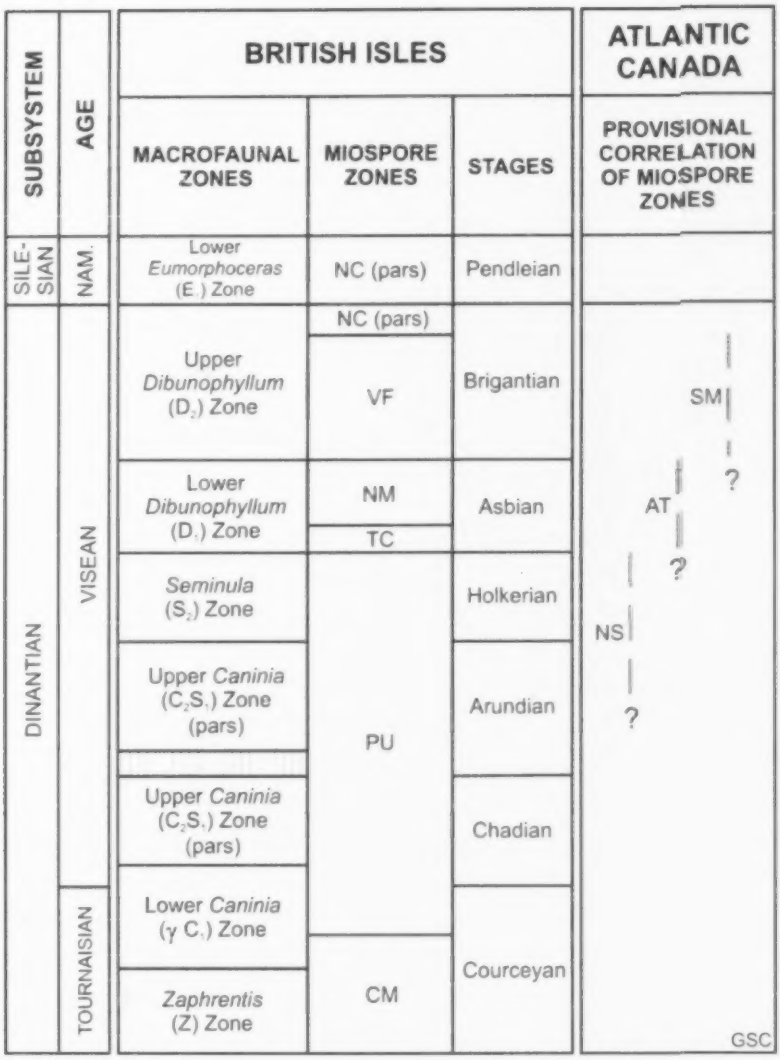

Figure 31. Provisional correlation of concurrent range zones in the British Isles and Atlantic Canada (after Utting et al. 1989). The boxes representing the macrofaunal and miospore zones and the stages are not drawn to scale. Red dotted lines indicate the possible range of miospore zones.

This formation comprises grey to black, parallel-laminated to thin-bedded wackestone and packstone (Fig. 33). The limestones are organic-bearing and commonly emit a petroliferous odor when freshly broken. Lithologs from the McCully Field report that the Macumber limestones are in part dolomitic and are commonly interbedded with tens of metres of anhydrite similar to that in the Upperton Formation. Thickness estimates for these two formations vary according to where their mutual contact is placed.

Finely laminated limestones of the Macumber Formation are interpreted as the deeper water (anoxic), lateral equivalents of the reef facies of the Gays River Formation (McCutcheon 1981). The Macumber Formation is inferred to be late early to middle Visean age on the basis of its stratigraphic position and interpreted lateral relationship with the spore-dated Gays River Formation.

Figure 32. Flaggy grey limestone of the Macumber Formation lying in parallel disposition above reddish brown conglomerate of the Hillsborough Formation. Outcrop is about 3.5 km south of Hillsborough (Fig. 2). Hammer for scale.

Figure 33. Limestone of the Macumber Formation, showing thin wackestone and packstone laminae and discontinuous, dark grey, organic-rich laminae. Diameter of coin is 2.7 cm.

Parleeville Formation

McCutcheon (1981) introduced the name Parleeville Formation for a heterogeneous suite of interbedded carbonate and clastic rocks that defines the basal marine unit of the Windsor Group in the Cocagne Subbasin from Parleeville to McGregor Brook (Fig. 4). In this area, the Parleeville Formation ranges in thickness from 1 m to 22 m and is the only exposed unit of the Windsor Group.

The relative proportion of carbonate and clastic beds in the Parleeville Formation differs from one location to another, but overall the formation contains grey to black algal boundstone, grey to reddish grey wackestone, packstone, sandstone, and mudstone. The strata locally contain brachiopods, pelecypods, gastropods, bryozoans, foraminifera, and ostracods, species similar to those observed in the Gays River Formation (McCutcheon 1981). The Parleeville Formation is ascribed an intertidal to sublittoral depositional environment on the basis of its faunal content, siliciclastic beds, and widely visible early diagenetic oxidation (red colour).

Two spore assemblages recovered from the Parleeville Formation by S. McCutcheon and studied by Barss (1980) were assigned a middle Visean (NS miospore zone) age (Fig. 31). A sample collected from the Parleeville Formation by the present author and studied by Dolby (1995a) also yielded an assemblage of NS miospore zone age.

Upperton Formation

The type section of the Upperton Formation appears in outcrops near Upperton (Fig. 2), 28 km southwest of Sussex (Anderle et al. 1979; McCutcheon 1981). Surface distribution of this formation has had to be largely inferred from karst topography, since the strata consist mainly of readily eroded anhydrite and gypsum, and outcrops are uncommon.

The Upperton Formation occurs above the Macumber Formation at Intervale in the IMDZ, and in the Smith Creek and Sussex areas in the Moncton Subbasin. In different surface sections, its thickness ranges from 40 m to 80 m. At the McCully Field, the drilled thicknesses in nine exploration boreholes vary from 42 m to 109 m. The true stratigraphic thicknesses are somewhat less than this; the thicker sections possibly could be repeated or partially repeated by faults (Wilson et al. 2006).

The Upperton Formation is composed primarily of mostly replacive and displacive anhydrite. The host rock of the anhydrite commonly is laminated limestone and less commonly is greenish grey mudstone. In exploration boreholes at the McCully Field, it often is difficult to define a boundary in the broad, gradational interval between limestone of the Macumber Formation and anhydrite of the Upperton Formation. In surface exposures, the Upperton

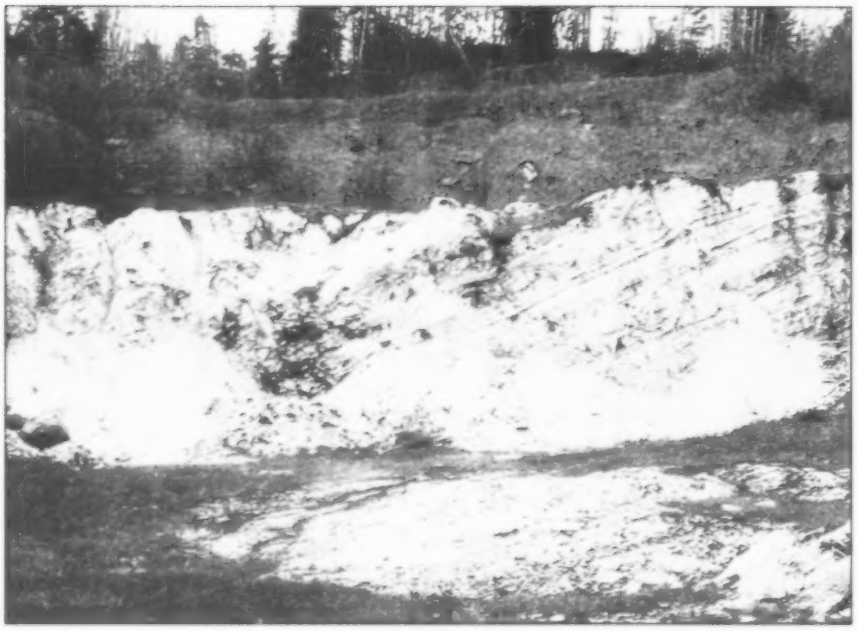

Figure 34. Quarry in gypsum of the Upperton Formation near Hillsborough (Fig. 2). Quarry face is about 8 m high.

anhydrite tends to be largely gypsified and, in places such as the Hillsborough area (Fig. 2), has been extensively quarried (Fig. 34). A subaqueous, deep-basin depositional environment has been interpreted for the Upperton Formation (McCutcheon 1981).

McCutcheon (1981) and McCutcheon (in Williams et al. 1985) reported that spores from the Upperton Formation in a borehole at Peekaboo Corner, southwest of Sussex, were of middle Visean age. Spores from stratigraphically equivalent beds from a borehole near Stewiacke, Nova Scotia, also indicated a middle Visean age (Utting 1980).

Cassidy Lake Formation

The Cassidy Lake Formation is in part conformably underlain by, and in part laterally equivalent to, the Upperton Formation (Anderle et al. 1979; McCutcheon 1981). The type section of the Cassidy Lake Formation occurs in borehole intersections at the former Clover Hill Mine (potash) 20 km southwest of Sussex (Anderle et al. 1979; McCutcheon 1981).

The Cassidy Lake Formation in the Penobsquis 'salt ridge' ranges in thickness from near-zero to approximately 400 m (Roulston et al. 1995) as a result of both salt doming or halokinesis (see Hinds and St. Peter 2006, this volume, their Fig. 7) and later structural deformation (Wilson 2005; Wilson et al. 2006). In the subsurface near Penobsquis (Fig. 4), rock salt and potash ore (sylvinite) are mined from the Cassidy Lake Formation by the Potash Corporation of Saskatchewan (PCS) Inc.

The Cassidy Lake Formation is composed of mostly clear, smoky, and brown halite, lesser brick-red sylvinite, and minor anhydrite with rare thin beds and laminae of grey mudstone and claystone (Fig. 35; Kingston and Dickie 1979; Roulston et al. 1995; Wilson et al. 2006).

Figure 35. Close-up view of the Cassidy Lake Formation, exposed in stope wall at the Potash Corporation of Saskatchewan (PCS) Inc. potash mine near Penobsquis (Fig. 4). Photograph shows halokinetically or tectonically folded potash ore with laminae of pink sylvinite, smoky grey halite, and dark grey clays and/or anhydrite. Vertical ridges are corrugation marks from the Alpine Miner.

These evaporite deposits have been described as deep-water, deep-basin deposits by McCutcheon (1981) on the basis on their internal stratigraphy and bounding relationships with the underlying and adjacent Upperton Formation and overlying Clover Hill Formation.

Borehole intersections indicate that the contact between the Cassidy Lake and Clover Hill formations is conformable. Where the Clover Hill Formation has been suberoded, such as on the flanks or over the crest of the Penobsquis salt ridge, the Cassidy Lake Formation has been partially or entirely removed. Suberosion typically is indicated by a zone of solution-collapse breccia comprising fragments of anhydrite and mudstone in a red muddy matrix. At such locations, the Cassidy Lake Formation is disconformably succeeded by the Mabou Group (Wilson 2005).

Spores reported by McCutcheon (in Williams et al. 1985) from Cassidy Lake evaporites were common to the NS miospore zone (Fig. 31; Utting 1987) and indicated a middle Visean age.

Clover Hill Formation

The type section of the Clover Hill Formation is defined from borehole intersections at the former Clover Hill potash deposit (Anderle et al. 1979; McCutcheon 1981). It appears only in the subsurface above the Cassidy Lake Formation in proximity to the Penobsquis salt ridge, where it was intersected in boreholes sunk during potash exploration in the 1970s and 1980s, and during natural gas exploration since 2000. The formation occurs in some boreholes but not others. Where present, it ranges in thickness from 2 m to 55 m and has an average thickness of about 15 m. Its absence is the result of either nondeposition or, more likely, suberosion.

Anderle et al. (1979) and McCutcheon (1981) described the Clover Hill Formation at Penobsquis as comprising a lower anhydrite interval, a medial halite interval, and an upper grey claystone interval. Their reported anhydrite-salt-claystone stratigraphy is somewhat idealized, since some boreholes intersect mostly grey claystone with only minor halite and very rare anhydrite. Although the Clover Hill evaporites are clearly marine deposits, the gradational contact (Anderle et al. 1979) of this formation with the overlying terrestrial redbeds of the Mabou Group implies a shallowing-upward sequence (McCutcheon 1981).

As with the Cassidy Lake Formation, the Clover Hill Formation has been suberoded in places over the crest of, and adjacent to, the Penobsquis salt ridge. Where suberosion has occurred, the stratigraphic position of the Clover Hill Formation is presently occupied by a zone of solution-collapse breccia comprising angular fragments of anhydrite and mudstone in a red muddy matrix (Wilson 2005). (In PCS mine terminology, this interval is called caprock.) The anhydrite and mudstone clasts quite likely are derived from dissolved beds of the Clover Hill Formation.

The conformable contact between the Clover Hill Formation of the Windsor Group and overlying fine-grained redbeds of the Mabou Group can be observed in some boreholes at Penobsquis. The contact is also inferred from seismic reflection profiles over the McCully Field (Hinds and St. Peter 2006, this volume). This conformable contact becomes an unconformity south of the McCully Field where the Mabou Group cuts across the underlying stratigraphy, and at the southern edge of the Moncton Subbasin where it rests directly on crystalline basement (Hinds and St. Peter 2006, this volume).

The Samp Hill beds in the Cocagne Subbasin is a name applied by McCutcheon (1981) to an 8 m thick interval of interbedded wackestones, mudstones, and fine-grained sandstones that lies above the Gays River Formation in the Havelock and Springhill area (Fig. 4). The basal 3 m of this unit is characterized by a chaotic intraformational conglomerate comprising wackestone clasts in a mudstone and sandstone matrix. McCutcheon (1981) interpreted the basal chaotic interval as representing a solution-collapse zone that resulted from the dissolution of evaporites that lay above the Gays River Formation. His assessment of this zone as an interval representing lost evaporite stratigraphy is reinforced by the fact that the Samp Hill beds are conformably overlain by about 4 m of 'transitional beds' consisting of green, fine-grained sandstone (McCutcheon 1981).

At Samp Hill, strata of the Clover Hill Formation are gradationally succeeded by fine-grained redbeds of the Mabou Group. This apparently conformable contact between the Windsor and Mabou groups would limit the entire Windsor Group to a total thickness of 45 m, wherein the Gays River Formation is 33 m thick, the Samp Hill beds are 8 m thick, and the transitional beds are 4 m (McCutcheon 1981). However, the complete time interval of the Windsor Group being represented by a mere 45 m of section in this location seems highly

unlikely and would imply the presence of a significant intra-Windsor hiatus. It therefore is recommended herein that the terms 'Samp Hill beds' and 'transitional beds' be abandoned. In this report, these beds are included in the Clover Hill Formation in the manner reported by Anderle et al. (1979).

A spore sample collected from the Samp Hill beds by McCutcheon (1981, his map 9) and reported by Barss (1979) yielded a middle Visean (NS miospore zone) age (Fig. 31). Spore collections from seven samples of grey siltstones and mudstones from grey transitional beds of the Clover Hill Formation in a potash exploration borehole south of Millstream area (Fig. 4) were assigned by Utting (1987) to his NS miospore zone.

Mabou Group

The type section for the Mabou Group occurs along Mabou River on Cape Breton Island, Nova Scotia (Norman 1935). In New Brunswick, stratigraphically equivalent clastic rocks lie conformably or unconformably above the Windsor Group and below the Boss Point Formation (Cumberland Group) and historically were called the Hopewell Group (Norman 1941a; Gussow 1953). Following the 1990s Federal/Provincial NATMAP project on the Maritimes Basin in southeastern New Brunswick, post-Windsor Group nomenclature was consolidated, and the name Mabou Group was applied to both the New Brunswick and Nova Scotia beds. For more on this subject, see Johnson and St. Peter (1997), and St. Peter and Johnson (1997).

In the present report area, strata of the Mabou Group are exposed in the Petitcodiac and McCully areas of the Moncton Subbasin, and above the Windsor Group in the Havelock and Parleeville areas of the Cocagne Subbasin (Fig. 4, 5). A northeast-trending tract of Mabou strata also lies on crystalline rocks of the New Brunswick Platform on the northwest side of the Belleisle Fault from Belleisle Creek to Salem.

The thickness of the Mabou Group in southern New Brunswick varies greatly. Some of the boreholes in the McCully Field (Fig. 4) in the Moncton Subbasin intersected more than 800 m of Mabou beds. In the Cumberland Subbasin, Mabou strata reach up to 2 km thick (Gussow 1953; Ryan and Boehner 1994), but on the New Brunswick Platform such as near Salem and Marrtown (Fig. 4), they are absent or measure only a few metres or tens of metres.

In the Sackville and Cumberland subbasins of southeastern New Brunswick (Fig. 2), Norman (1941a) subdivided the Mabou Group into (in ascending order) the fine-grained Maringouin Formation, the sandy Shepody Formation, and the conglomeratic Enragé Formation. A cursory examination of selected borehole lithologs from potash exploration holes south of Lower Millstream and natural gas wells in the McCully Field (Fig. 4) shows

that the Mabou Group in the Cocagne and Moncton subbasins is also a coarsening-upward sequence. Reddish brown mudstones dominate the lower part, pass progressively upsection into more sandstone, and are capped by massive, thick-bedded, polymictic conglomerates. The well logs also indicate that the lower fine-grained interval increases remarkably in thickness from about 75 m in the southwestern part of the Lower Millstream area in the Cocagne Subbasin to greater than 800 m in the northeastern borehole at the McCully Field in the Moncton Subbasin. This northeasterly thickening of the fine-grained redbeds is concomitant with a northeasterly thinning of the coarse upper part of the Mabou sequence.

Farther northeast, in the Petitcodiac region of the Moncton Subbasin, the Mabou Group has not been examined in any detail. Where rocks of the group lie above sulphates of the Windsor Group in that area, McCutcheon (1981) described the Mabou redbeds as generally comprising a fine-grained facies of sandstone and mudstone with lesser conglomerate. The dominance of fine-grained strata at Petitcodiac is consistent with its on-trend relationship to the thick, fine-grained lower part of the Mabou Group in the McCully Field area.

On the New Brunswick Platform near Marrtown and Pearsonville where the Mabou section is very thin (Fig. 4), the strata are composed nearly entirely of red to locally grey polymictic conglomerates, red to grey lithic sandstones, and very minor reddish brown mudstone and calcrete. St. Peter (1992) and Johnson (2003) correlated these beds with the Shin Formation, which has its type section about 80 km southwest of Pearsonville (Fig. 4) on the south side of the New Brunswick Platform (van de Poll 1967). The Shin Formation is exposed fairly continuously from its type section eastward to Pearsonville. Its stratigraphic position is confirmed locally by the fact that it overlies fossil-bearing marine carbonates of the Windsor Group (MacKenzie 1964) and underlies clastic beds of the Cumberland Group (St. Peter 1992; McLeod et al. 1994a, 1994b; Johnson 2003).

Coarse clastic rocks exposed in the northeast-trending Belleisle Creek–Springhill tract (Fig. 4) previously were assigned to the Moncton Group (that is, pre-Windsor Group) by Alcock and MacKenzie (1946) and to the Memramcook Formation by Gussow (1953) and all subsequent workers. This assignment was based largely on the fact that the conglomerates and sandstones lie on or against crystalline rocks of the New Brunswick Platform from Belleisle Creek to Harry Brook. In the absence of any biostratigraphic control, it seemed reasonable to assume these beds represented the oldest strata of the Maritimes Basin in the area. However, recent mapping in New Brunswick has demonstrated the potential fallacy of this reasoning: palynologically dated Mabou strata, for example, are now known to lie directly on the crystalline Caledonia Uplift in the Cumberland Subbasin southwest and northeast of Alma (Fig. 2; St. Peter 1997a; St. Peter and Johnson 1997).

The coarse redbeds in the Belleisle–Springhill tract have been assigned to the Mabou Group for the following reasons. They are juxtaposed against basement rocks along the

Kierstead Mountain Fault on the northwest, and against the Bloomfield Formation via the Belleisle Fault on the southeast. Farther northeast, along Highway 2, the same beds occur on either side of the Belleisle Fault. At this locale, they lie unconformably on the Bloomfield Formation (Horton Group) on their southeast side and are unconformably overlapped by the Minto Formation (Pictou Group) on their northwest side. Still farther northeast, just west of Springhill, these conglomerate beds dip off a previously unknown inlier of Windsor limestone (Gays River Formation), thus confirming by their stratigraphic position that they belong to the Mabou Group (Fig. 4). Given the foregoing, the earlier location of the Kierstead Mountain Fault in the Harry Brook area by St. Peter (1992) is erroneous. Consequently, the distinction made between the Shin Formation (Mabou Group) on the northwest side of that fault and the Memramcook Formation (Horton Group) on the southeast side has been reassessed. Beds on either side of this interpreted fault are lithologically indistinguishable and therefore are herein assigned to the Mabou Group.

The Mabou rocks in the Belleisle Creek–Springhill tract are characterized by a drab maroon colour. Most outcrops are dominated by polymictic pebble to cobble conglomerate in medium to very thick beds that typically are massive and crudely interstratified with medium- to very coarse-grained lithic sandstones (Fig. 36). Beds of reddish brown siltstone and mudstone are present locally. The conglomerates in places have clasts up to boulder size in outcrops adjacent to the basement. The clasts range from angular to well rounded and comprise mainly dark-coloured metasedimentary, volcanic, and plutonic rocks; vein quartz clasts also are locally seen. These clasts are comparable to lithotypes in the basement situated immediately northwest. The Mabou conglomerates can be distinguished from adjacent redbed units by their darker coloured matrix and clasts, and also by their pervasive desert varnish on pebbles and cobbles, particularly on the more rounded clasts.

The Mabou Group is interpreted herein to represent an alluvial complex sourced from the north and west. This is suggested by the presence of 1) abundant, coarse-grained lithotypes on the New Brunswick Platform and in the southwestern part of the Cocagne Subbasin in the Lower Millstream area, in conjunction with 2) the mainly fine-grained lithotypes (particularly in the lower part of the sequence) in the eastern part of the Moncton Subbasin near Petitcodiac. The interpretation is also supported in part by the fact that the Mabou section is several hundreds of metres thick southeast of the Belleisle and Kierstead Mountain faults, but absent or only a few tens of metres thick on the New Brunswick Platform. These contrasting thicknesses and lithotype distributions imply that the Belleisle Fault system approximates a Late Mississippian depositional escarpment and that the sequence on the New Brunswick Platform represents a condensed section equivalent to the thicker Mabou basinal sequence farther east.

The lower contact of the Mabou Group with the underlying Windsor Group and older rocks has already been described in detail (see *Cassidy Lake Formation* and *Clover Hill*

Figure 36. Typical conglomerate of the Mabou Group exposed in the tract from Belleisle Creek to Springhill (Fig. 4). Photograph shows poorly sorted and angular to subangular clasts of volcanic and metasedimentary rocks with minor quartz and fine-grained plutonic rocks. Note the thin, lenticular bed of sandstone behind hammer head.

Formation). The upper boundary of the Mabou Group is a profound unconformity, as implied by the lithological contrast between the compositionally immature and locally derived Mabou strata, and the compositionally and texturally much more mature strata of the overlying Cumberland Group (see *Boss Point Formation*, below). The contrast suggests an extensive period of local erosion and the establishment of a broad new base level.

The Mabou–Cumberland unconformity is clearly demonstrated on the New Brunswick Platform in the Pearsonville and Marrtown areas. Here, the grey Cumberland strata overlap the Mabou redbeds and in part lie on crystalline basement, implying that the Mabou sections represent the erosional remains of valley-fill sequences lying on the old basement topography. Additional evidence of the unconformity at the close of the Mabou depositional cycle is shown in the Moncton Subbasin near Mount Pisgah (Fig. 4), where both Windsor and Mabou groups are overlapped by the Cumberland Group. The extent of the erosion is also evident beyond the report area, as exemplified by an unconformity at the base of the Cumberland Group along the southern margin of the Moncton Subbasin near Parkindale (Fig. 2). In that location, the Boss Point Formation directly lies on the Albert Formation (Fig. 3, 37).

No grey (unoxidized) fine-grained Mabou beds have been located in the report area that could offer the potential to yield spores. However, spore assemblages collected from beds in the eastern Moncton Subbasin, Sackville Subbasin, and western Cumberland Subbasin indicated a late Visean to early Namurian age for the Mabou Group (St. Peter and Johnson 2005). These same palynological data recorded a middle Namurian time break at the MabouCumberland boundary, which strongly supports the lithological evidence of the unconformity described above.

Figure 37. Unconformity near Parkindale (Fig. 2) between dipping grey mudstone of the Albert Formation and gently inclined quartz-pebble conglomerate and flaggy sandstone of the Boss Point Formation. Dashed line highlights bedding in the Albert Formation.

Cumberland Group

Boss Point Formation

The type section of the Cumberland Group including that of its basal unit, the Boss Point Formation, occurs in the Cumberland Subbasin of Nova Scotia on the seacliffs of Chignecto Bay (Bell 1944; Ryan et al. 1991). The Boss Point Formation in the report area occurs on the New Brunswick Platform in the Pearsonville–Goshen area (Fig. 4). Boss Point strata are also found in the Moncton Subbasin northeast of Sussex in two northeast-trending synclinal structures. One lies southeast of Springdale, and the other occurs in the Mount Pisgah–Dunsinane area (Fig. 4, 5; Gussow 1953).

Johnson (2003) was the first to distinguish the Boss Point Formation from the overlying Minto Formation (Pictou Group) in the Pearsonville–Goshen area. The Boss Point Formation there comprises yellow-brown-weathering, grey, quartz-pebble conglomerate; quartz-pebble sandstone; and fine- to medium-grained sandstone with lesser interbeds of grey to olive, very fine-grained rippled sandstone, siltstone, and mudstone. Coalified plant fragments are locally common.

Figure 38. Outcrop of the Boss Point Formation, showing typical trough cross-beds in coarse-grained quartzose arenite.

Lithotypes of the Boss Point Formation in the Cumberland Subbasin and eastern Moncton Subbasin generally are arranged into multiple-stacked, fining-upward sequences (St. Peter 1997a; St. Peter and Johnson 1997). The compositionally mature conglomerate and coarser sandstones typically are trough or planar cross-stratified (Fig. 38). Parallel or flat beds with parting lineation are commonly observed in the medium-grained beds in the medial parts of fining-upward sequences.

In his work on the Boss Point Formation in southeastern New Brunswick, Browne (1990) proposed a braided-stream channel environment for the conglomerates and sandstones, and various interfluvial environments (crevasse splay, lacustrine, and lacustrine-deltaic) for the finer grained sandstones and mudrocks. The presence of local calcrete layers and rooted horizons attests to paleosol development on the desiccated interchannel tracts.

The Boss Point Formation is overlain unconformably by the Pictou Group (Minto and Salisbury formations) in the report area. Near Goshen (Fig. 4), the Boss Point Formation thins out entirely and, along with the adjacent basement rocks, is overlapped by the Minto Formation. Farther northeast, the Minto Formation lies directly on redbeds of the Mabou Group. An analogous relationship exists in the Petitcodiac area in the Moncton Subbasin, where the Boss Point Formation is absent and the late Westphalian Salisbury Formation lies on the early Namurian (?) Mabou Group (Fig. 5). A rich palynological data set exists for the Boss Point Formation in southeastern New Brunswick (St. Peter and Johnson 2005) and has indicated an age ranging from late Namurian to early Westphalian B.

Pictou Group

The Pictou Group represents the youngest sequence of strata in the Maritimes Basin. It is widely distributed, relatively thin, and oversteps all older units of the Maritimes Basin and adjoining basement rocks. The type section of this group occurs in Pictou County, Nova Scotia, on the West Branch River John from the mouth of the river upstream to Mine Brook (Bell 1944).

In New Brunswick (St. Peter 1997a; St. Peter and Johnson 1997), as in the type section, the Pictou Group unconformably overlies the Boss Point Formation. The Pictou Group tilts regionally toward the Gulf of St. Lawrence (Fig. 2) but otherwise is only gently warped into broad, northeast-trending folds with wavelengths of several kilometres (Gussow 1953). In some late fault zones, however, the Pictou strata are steeply inclined.

In the report area, the Pictou Group comprises the Salisbury, Minto, and Richibucto formations. The distribution, sedimentology, and internal stratigraphy of these formations have little bearing on the geological history of the Cocagne Subbasin and IMDZ and consequently are discussed below only briefly.

Salisbury Formation

The Salisbury Formation was introduced by Gussow (1953) to describe the basal unit of the Pictou Group in the Moncton Subbasin. The formation comprises mostly red mottled mudstone, siltstone, and greyish maroon, very fine-grained sandstone. It also features lesser interbeds of pinkish grey, red and grey, fine- to medium-grained, pebbly, lithic and feldspathic arenite; quartz arenite; intraformational mudstone-clast conglomerate; polymictic pebble conglomerate; and very rare thin coal seams (St. Peter and Johnson 1997).

On the southeast side of the Berry Mills Fault, the Salisbury Formation extends from its type area near Salisbury southwestward to Dunsinane and northeastward as far as Moncton (Fig. 4, 5). On the northwest side of the Berry Mills Fault immediately west of Salisbury, it unconformably overlies the Sussex Group. The unconformable basal contact of the Salisbury Formation at this location is clearly evidenced by spore-dated (Westphalian) Salisbury beds overstepping the late Tournaisian Weldon and Briggs Cross formations of the Sussex Group (Fig. 5). The upper contact of the Salisbury Formation with the overlying Richibucto Formation is not exposed in New Brunswick (Johnson 1996). However, the apparent absence of any structural discordance across the boundary, as determined from both outcrop bedding data and seismic reflection profiles, suggests that the upper contact is conformable.

Numerous spore assemblages from the Salisbury Formation yielded late Westphalian B and Westphalian C dates (St. Peter and Johnson 2005). The upper age range is consistent with the inferred conformable relationship with the overlying Richibucto Formation.

Minto Formation

The Minto Formation was introduced by Muller (1951) to represent a sequence of buff to grey sandstones, pebbly sandstones, quartz-pebble conglomerates, minor grey siltstone and shale, and a thin, mineable coal seam in the Minto–Chipman area of central New Brunswick (Fig. 2). The formation is about 85 m thick at its type section on Newcastle Creek near Minto. St. Peter (2000) expanded the definition of the Minto Formation to incorporate the overlying Hurley Creek and Sunbury Creek formations as defined by Muller (1951). This expanded application is used herein (see Johnson and St. Peter 1997).

The Minto Formation in the report area comprises exposed beds on the New Brunswick Platform that extend from Marrtown and Goshen in the southwest to north of Monteagle in the northeast (Fig. 4, 5). The beds generally are similar to those described by Muller (1951) for the Minto, Hurley Creek, and Sunbury Creek formations in the Minto area. They comprise repetitively stacked, fining-upward sequences characterized by basal grey to red, sandstone-dominated intervals and upper mainly red mudstone-dominated intervals (Hamilton 1960, 1961, 1962a, 1962b). The intervals range in thickness from a few metres to tens of metres. In general, sandstones of the Minto Formation tend to contain somewhat more quartz and less feldspar and mica than do overlying sandstones of the Richibucto Formation.

The Salisbury and Minto formations are both basal units of the Pictou Group and are approximately equivalent in age. A preliminary assessment of available data suggests the Minto Formation lies on the western part of the New Brunswick Platform, while the finer grained Salisbury Formation occupies the eastern parts of the New Brunswick Platform and the Moncton Subbasin.

The basal contact of the Minto Formation, like that of the Salisbury Formation, is an unconformity. On the south side of the New Brunswick Platform (Fig. 4, 5), Minto strata overstep the older Mabou, Windsor, and Sussex groups west of Monteagle, and the Cumberland Group and basement rocks near Goshen. The contact between the Minto Formation and basement rocks is exposed along Ridge Brook just north of Havelock (Fig. 4). Westphalian C spores (Ball et al. 1981) have been obtained from that location, which should preclude any misguided assignment of those beds to the late Namurian–early Westphalian Boss Point Formation (Cumberland Group).

The regional unconformity at the base of the Salisbury and Minto formations is the key reason for retaining this boundary as the base of the Pictou Group in New Brunswick. No other clear depositional or chronological break has yet been demonstrated in younger Westphalian or Stephanian strata in New Brunswick (St. Peter and Johnson 2005). Thus, no geological reason exists to move the basal boundary of the Pictou Group upsection to a transitional grey-to-red colour boundary, as has been recommended for Nova Scotia (Ryan et al. 1991).

Numerous spore assemblages from the Minto Formation (see compilation in Ball et al. 1981) and unpublished collections by the present author have indicated a late Westphalian B to Westphalian C age for the formation. Plant assemblages from the Minto coal seam support a Westphalian C age (Muller 1951).

Richibucto Formation

Gussow (1953) introduced the Richibucto Formation to designate beds of the Pictou Group near the town of Richibucto (Fig. 2, inset). He estimated the thickness of the Richibucto Formation at about 300 m on the Westmorland Uplift west of Moncton and toward the Northumberland Strait (Fig. 2), where the formation is conformably succeeded by the Tormentine Formation (Johnson 1996). Gussow's definition was expanded by Carr (1968), who included underlying beds of Gussow's Scoudouc Formation in a revised Richibucto Formation, thus effectively abandoning the name Scoudouc Formation.

Within the report area, the Richibucto Formation overlies the Salisbury Formation north of Moncton and is in contact with the Briggs Cross Formation (Sussex Group) along the Berry Mills Fault from Lutes Mountain to Irishtown (Fig. 5, 6). Northeast of Irishtown, the Richibucto Formation lies on the northwest side of the Berry Mills Fault and unconformably onlaps the Horton and Sussex groups on the north end of the IMDZ. On the northwest side of the Smith Creek Fault, the distribution of the Richibucto Formation and its contact with the underlying Salisbury Formation is less certain, as no rigorous investigations have yet been conducted to determine the lithological boundary in that area.

The Richibucto Formation typically comprises grey and brownish red, commonly micaceous, lithic and arkosic sandstone, pebbly sandstone, and intraformational mudstone-clast conglomerate. Interbeds and intervals of brownish red to brick-red and lesser grey siltstone and mudstone occur up to several metres thick. Sandstones of the Richibucto Formation in the report area have considerably more mica, feldspar, and lithic grains than do those of the Salisbury Formation (Johnson 1996).

Spore assemblages from several localities in the Richibucto Formation have yielded ages that range from late Westphalian C to Stephanian (St. Peter and Johnson 2005).

Structural Geology

Various papers published on the structural geology of the Maritimes Basin (see *Introduction*) have implications for interpreting the stratigraphy of southeastern New Brunswick. A review of these papers reveals an emphasis on the importance of faults, particularly major faults, especially as to whether the fundamental movement on those faults was compressional (Gussow 1953), lateral (Webb 1963), extensional (Belt 1968; Ruitenberg and McCutcheon 1982), or transtensional–transpressional (Leger and Williams 1986; St. Peter 1993; Nickerson 1994). Two significant factors have contributed to confusion over the deformation style and displacement sense along those faults, namely 1) an inadequate chronological (biostratigraphic) control on lithostratigraphic units, and 2) an insufficient understanding of the presence and position of unconformities in the stratigraphic record.

The following discussion of major and minor faults relies on new palynostratigraphic and lithostratigraphic evidence to infer constraints on the tectonic history of the Cocagne Subbasin and IMDZ. Major faults in the report area are represented by the Belleisle Fault, North River Fault, and Jordan Mountain Fault (Fig. 4, 5, 6). These faults have had a significant, demonstrable influence on the deposition and location of sedimentation. Minor faults in the report area include the Smith Creek Fault, Springhill and Cosman Settlement faults, Berry Mills and Kennebecasis faults, and Penobsquis Fault. These minor faults may have given rise to substantial stratal offset but appear to have had no influence on sedimentation. As such, they show no record of movement during deposition and thus are not basin-bounding structures.

Major Faults

Belleisle Fault

The stratigraphic record indicates that the Belleisle Fault, together with its associated splay, the Kierstead Mountain Fault, is a fundamentally important structure. This fault system had a controlling influence on the margin of the Cocagne Subbasin during deposition of the Horton, Sussex, and Mabou groups (Fig. 4, 5). During clastic depositional cycles in the Late Devonian and Early Carboniferous, the crystalline basement on the northwestern side of the fault (that is, the New Brunswick Platform) acted as a paleohigh and source area. This exhumed basement area remains visible today (Fig. 39).

Columbia boreholes in the Monteagle area (Fig. 5) provide the best evidence for interpreting the influence of the Belleisle Fault on sedimentation of the Horton and Sussex groups. Borehole J-67 intersected 14 m of Minto Formation immediately below surface and continued to a total depth of 2499 m, passing through massive polymictic conglomerates

Figure 39. View looking west across the Cocagne Subbasin from the west side of Jordan Mountain. Photograph shows the Belleisle Fault trace (B) and Kierstead Mountain Fault trace (K). Crystalline basement of the New Brunswick Platform forms the highlands along the horizon.

with only minor interbeds of lithic sandstones. The sequence contained no grey mudstones that potentially could yield a spore assemblage.

In contrast, borehole G-56 situated 3 km south was drilled to 1500 m and intersected four distinct units. From top to bottom the units were 1) 642 m of polymictic conglomerates and lithic sandstones, 2) 333 m of red and minor grey siltstone, shales, and rare black shale, 3) 284 m of fining-upward, stacked cycles of mostly red sandstone, siltstone, and shale, and 4) 241 m of polymictic conglomerates with lesser interbeds of sandstone and mudstone. Unit 2 yielded a spore assemblage of Tournaisian Zone 2B age, and unit 3 provided a sample of Tournaisian Zone 1 age (Fig. 10; Dolby 2004). On the basis of surface geology and palynological data, these four stratigraphic units were interpreted as being 1) Ridge Brook Formation, 2) upper McQuade Brook Formation or lower Albert Formation, 3) McQuade Brook Formation, and 4) Memramcook Formation, respectively.

Correlation of the strata in boreholes J-67 and G-56 leads to the inescapable conclusion that units 2 and 3 in G-56 do not occur in J-67, and that the 2485 m of conglomerates in J-67

must represent the Ridge Brook Formation in the upper part of the hole and probably the Memramcook Formation in the lower part of the hole. The implication is that the Ridge Brook Formation represents the entire Sussex Group at J-67, has 'cut out' upper formations of the Horton Group, and lies unconformably on the Memramcook Formation at some unknown depth in the borehole. Conglomerates of the Horton and Sussex groups in borehole J-67 are greater than 2 km thick, indicating the proximity of this area to a major active escarpment, that being the Belleisle Fault. At the type section of the Ridge Brook Formation southwest of borehole J-67, cross-beds and clast types confirm an adjacent northern source for sedimentary detritus of the Sussex Group.

The Springhill Fault is situated subparallel to, and a few kilometres south of, the Belleisle Fault in the Monteagle area (Fig. 5). It is unknown what, if any, role the Springhill Fault played in causing the contrast of strata in boreholes J-67 and G-56 (Fig. 5). West of Monteagle, the Ridge Brook Formation on the south side of the fault is juxtaposed against the Windsor and Mabou groups on the north side, implying that the south side has risen. Northeasterly along the fault, Ridge Brook strata are in contact with beds of the Pictou Group, which implies a Late Carboniferous or younger up-on-the-south displacement. This Late Carboniferous south-over-north movement on the Springhill Fault, however, does not explain why borehole G-56 contains more apparent Horton stratigraphy preserved beneath the basal Sussex unconformity than does borehole J-67. For example, if the Springhill Fault existed between the sites of boreholes J-67 and G-56 before the Sussex depositional cycle began, the fault must have had a down-on-the-south displacement to explain the absence (due to erosion?) of the McQuade Brook and Albert formations in borehole J-67. Other faults in the Maritimes Basin of southern New Brunswick also demonstrate contrasting senses of displacement at different times during their movement history (Park and St. Peter 2005).

On the west side of the basement block at Jordan Mountain in the Cocagne Subbasin, and between the Berwick and Belleisle faults, strata of the Albert Formation and overlying Bloomfield Formation (Horton Group) form a section about 1.5 km thick that dips about 30°NW toward the Belleisle Fault and adjacent basement of the New Brunswick Platform (Fig. 4). The presence of the Memramcook Formation adjacent to the Albert Formation on the southeast side of the Berwick Fault suggests that probably several hundred metres of Memramcook strata also underlie the Albert Formation on the northwest side of that fault.

Geometry and thickness of the northwesterly dipping section of the Horton Group between the Berwick and Belleisle faults strongly suggest that a significant part of what must have been a complementary southeasterly dipping Horton section is concealed beneath the Mabou Group bordering the New Brunswick Platform. It thus seems likely that crystalline rocks of the New Brunswick Platform became uplifted and overrode the Horton section along the Belleisle Fault system. Timing of that displacement is not clear. However, the absence of Horton strata beneath the unconformity at the base of the Mabou Group on the New

Brunswick Platform implies that most movement occurred before the Mabou depositional cycle began.

As noted earlier (see *Mabou Group*), the limited thickness of the Mabou section on the New Brunswick Platform, in contrast with its much greater thickness in the Cocagne and Moncton subbasins on the southeast side of the Belleisle and Kierstead Mountain faults, indicates that those faults defined an active escarpment or basin-bounding fault system during the Late Mississippian. The pattern of coarse and fine lithofacies distribution of the Mabou Group is consistent with this structural interpretation. It is unknown whether the Mabou-related fault surfaces dipped southeast and underwent dominantly down-on-the-southeast (extensional) movement or dipped northwest and underwent dominantly up-on-the-northwest (compressional) movement. Clasts within conglomerates of the Mabou Group are comparable to basement rock types of the adjacent New Brunswick Platform, suggesting that the Belleisle Fault system did not undergo significant lateral displacement. This conclusion was reached independently by Brown and Helmstaedt (1970).

The Belleisle Fault system had little or no influence on the thickness or distribution of lithofacies for either the Cumberland or Pictou groups, which suggests that the fault was dormant during Late Carboniferous sedimentation. Some evidence exists for later displacement on the Belleisle Fault. A seismic reflection profile (Irving/Chevron line 93) along Highway 126 near borehole 428 (Fig. 5), for instance, shows inclined reflectors in the Pictou Group. However, the amount of vertical offset following the Pictou depositional cycle is minimal, as implied by relative depths of the unconformity at the base of the Pictou Group on either side of the fault trace. The subhorizontal disposition of bedding in outcrops in close proximity to the interpreted surface trace of the fault supports the interpretation of such minimal displacement.

North River Fault

The North River Fault is a major break between two contrasting domains in the northeastern area of the IMDZ and can be traced from North River northeasterly to near the headwaters of Shediac River (Fig. 5, 6). The domain on the northwestern side of the fault comprises a very thick sequence of McQuade Brook, Albert, and Indian Mountain formations (Horton Group) that dips moderately to steeply southeast toward the fault. Rocks of the Albert and Indian Mountain formations adjacent to the fault trace are not proximal or upland facies. Consequently, the fault at this location does not represent a paleobasin margin for the Horton depositional cycle.

In contrast, the domain on the southeastern side of the fault is floored by basement rocks that are overlain by only minor erosional remnants of the Horton Group and an unconformably overlying thick sequence of the Sussex Group (Fig. 40). Coarse

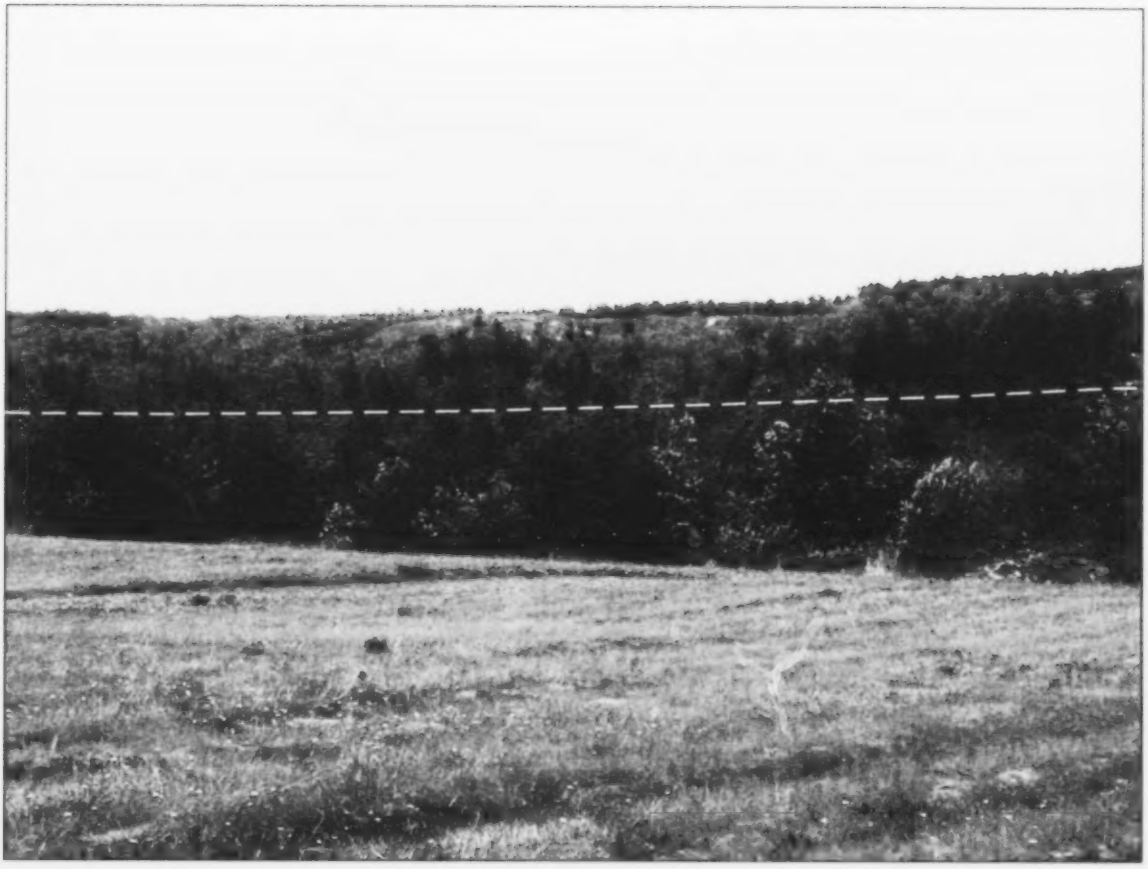

Figure 40. North River Fault trace (dashed line) viewed from Stilesville looking south (Fig. 6). The field in the foreground is underlain by the Albert Formation. The ridge in the background on the hanging-wall side of the fault is composed of conglomerate of the Stilesville Formation (Sussex Group), and granitoid basement. The Moncton Crushed Stone Quarry is visible at the top of the ridge.

megabreccia occurs in the Stilesville Formation at the base of the Sussex Group near the fault at The Gorge. The existence of the megabreccia, coupled with the regional distribution of the proximal Briggs Cross Formation and the distal southwesterly thickening Weldon Formation, strongly suggests that the North River Fault played a significant role in the deposition of Sussex strata in the Indian Mountain area.

The juxtaposition of the southeasterly dipping panel of Horton strata on the northwest side of the North River Fault against a basement block with Sussex strata on the southeast side of the fault implies a major loss of Horton strata on the southeastern side of the former Horton basin. (Whether this panel of Horton strata is part of the Cocagne or Moncton subbasin is addressed in *Conclusions*.) The basement block and its Sussex cover could have been thrust- or reverse-faulted northwestward over buried Horton strata in the subsurface.

The presence of The Gorge Fault, which is roughly parallel to (and immediately south of) the North River Fault, shows that northwestward-directed compression took place in this area (Fig. 6). For example, the southeastern or hanging-wall side of The Gorge Fault comprises

Figure 41. Exposure of the Gorge Fault, showing basement (B) on the right with a thin cover of megabreccia of the Stilesville Formation (S) juxtaposed against the Gautreau (G) and Briggs Cross (BC) formations on the left. The contact between the Gautreau and Briggs Cross formations is also a fault. Outcrop is situated on the east side of the abandoned Tri-Gil quarry at The Gorge (Fig. 6).

basement rocks with a cover of Sussex strata, whereas the northwestern, footwall side exposes only rocks of the Sussex Group (Fig. 41; St. Peter and Park 2006a, 2006b, 2006c). Displacement on the North River Fault, however, is not a simple case of northwestward-directed compression. A detailed study of better exposures adjacent to the fault has yielded kinematic evidence that the fault underwent significant right-lateral movement (Park, pers. comm.).

Near the headwaters of North River, the North River Fault is interpreted to be cut off by the Smith Creek Fault (Fig. 5, 6). It is unclear how the North River Fault traverses southwest of the cutoff, or if it continues to mark the boundary between a panel of Horton strata on the northwest and a basement block with a cover of Sussex strata on the southeast. Rocks of the Horton Group have been intersected at depth in all four Columbia boreholes (J-67, G-56, A-47, and L-18) in the Monteagle area (Fig. 5), which suggests that the structural boundary of the Horton basin lies south of those boreholes.

Jordan Mountain Fault

The Jordan Mountain Fault was first recognized and mapped as a northwest-dipping reverse structure by Gussow (1953). It can be traced from Searsville in the southwest to Jordan

Mountain in the northeast (Fig. 4). Primarily because outcrops in this area are sparse, it is unknown whether the fault continues northeast of the basement rocks at Jordan Mountain.

The fault trace is exposed in a roadside outcrop at Jordan Mountain, where about 10 m of black graphitic rhyolite are sheared against angular-clast conglomerate of the Ridge Brook Formation. The shear zone marking the Jordan Mountain Fault dips northwesterly in a mineral exploration hole on the southeast side of Jordan Mountain. The hole was drilled vertically in basement rocks and bottomed in conglomerates of the Ridge Brook Formation. This intersected shear zone implies that the basement and possibly the flanking rocks of the Horton Group on its northwest side were uplifted and overrode rocks of the Ridge Brook Formation (Sussex Group) to the southeast.

The basement rocks at Jordan Mountain and rocks of the Horton Group immediately northwest of them together represent a block that bounds the northwestern side of a thick section of Sussex strata to the southeast (Fig. 4). The Ridge Brook Formation on the southeast side of the Jordan Mountain basement is a fining-upward section that dips southeast. Clasts and cross-beds in conglomerates and sandstones of the Ridge Brook Formation indicate a southeasterly transport away from Jordan Mountain (that is, a northwestern source). A northwestern source of Ridge Brook detritus is also indicated east of Mount Middleton, where conglomerates contain recycled sandstone clasts that almost certainly are derived from Horton beds on the northwest side of the Jordan Mountain Fault.

The presence of exhumed basement on the northwest (hanging-wall) side of the Jordan Mountain Fault suggests that basement rocks may lie immediately beneath the Ridge Brook Formation on the southeast (footwall) side of the fault. For example, the Ridge Brook Formation is interpreted to lie unconformably on basement on the north side of Jordan Mountain (Fig. 4), which implies that the formation may lie directly on buried basement on the southeast side of the fault trace. However, Gussow (1953, his Fig. 9) argued that a thick section of Horton strata lies between basement rocks and the Ridge Brook Formation (his Hillsborough Formation) on the southeast side of the fault. His interpretation likely was influenced largely by the erronous belief that the Gautreau and Weldon formations, which occur on the southeast side of the Smith Creek Fault in the adjacent IMDZ, were units within the Horton Group.

The possibility that strata of the Sussex Group lie directly on basement southeast of the Jordan Mountain Fault (meaning that the Horton Group is absent) is further suggested by northwest–southeast-oriented seismic reflection profiles in the McCully Field, as interpreted by Wilson (2005). According to Wilson (2005), a thick section of the Horton Group occurs in the southern part of the field, lapping onto basement rocks along the northwestern side of the McCully Field. He also interpreted that rocks of both the basement and Horton Group are unconformably overlain by strata of the Sussex Group.

The interpreted seismic profiles at McCully and surface geology at Jordan Mountain suggest that no Horton section exists in the subsurface between the north side of the McCully Field and the exhumed basement at Jordan Mountain. Under that scenario, a buried crystalline ridge must be situated between sections of Horton strata in the Moncton Subbasin (including the McCully Field) and the Cocagne Subbasin northwest of Jordan Mountain. If a buried basement ridge does lie southeast of Jordan Mountain, and if it separates Horton strata of the Cocagne and Moncton subbasins, was that ridge present during the Horton depositional cycle and continuous with the crystalline rocks presently exposed at The Gorge? Or, did that ridge begin to rise only as a result of inversion following deposition of the Horton strata?

These questions cannot be answered, because it is not known if strata of the Sussex Group exposed in the IMDZ from Smith Creek to The Gorge overlie basement throughout the area, or if the zone also contains buried rocks of the Horton Group. If buried Horton strata do occur in the southern domain of the IMDZ between Smith Creek and The Gorge, the basement topography, by implication, was breached in the Early Mississippian, which would suggest that the Cocagne and Moncton subbasins (as herein defined) were a single Horton depocentre.

Minor Faults

Smith Creek Fault

The Smith Creek Fault (see Gussow 1953) is a significant northeast-trending structure that can be traced from Smith Creek in the southwest to McQuade Brook in the northeast, where it defines the northwestern boundary of the IMDZ (Fig. 2, 4, 5, 6). The break is not exposed in outcrop, but poor-quality Irving/Chevron seismic reflection profiles from the 1980s suggest that it is a steep, southeast-dipping structure. The Smith Creek Fault is interpreted to be largely compressional, with the southeast block overriding the northwest block. This sense of displacement is most obvious in the northeast, where the Lower Mississippian McQuade Brook Formation on the southeast is in contact with the Pennsylvanian (Westphalian) Pictou Group on the northwest (Fig. 5, 6).

Considerable early movement on the Smith Creek Fault almost certainly took place following the Sussex depositional cycle. For example, late Tournaisian proximal conglomerates and sandstones of the Ridge Brook Formation are juxtaposed against late Tournaisian distal mudstones, siltstones, and fine-grained sandstones of the Gautreau and Weldon formations. No surface evidence exists of an intermediate or transitional facies between proximal and distal sections of the Sussex Formation, which implies that considerable shortening must have occurred across their mutual boundary, the Smith Creek Fault. Evidence of that shortening is shown by the contrast between the shallowly dipping Ridge Brook Formation

on the northwest side of the fault, and the moderately to steeply dipping, fine-grained strata of the Gautreau and Weldon formations in the IMDZ on the southeast side of the fault.

Springhill and Cosman Settlement Faults

The Springhill Fault is a northeast-trending structure that crosses the Cocagne Subbasin and links the Belleisle Fault system with the Smith Creek Fault (Fig. 4, 5, 6). It was first identified by St. Peter (1995h) along Ridge Brook and near Springhill, where the normally shallowly dipping beds of the Ridge Brook and Gays River formations are thrown into moderately steep dips. As with the Smith Creek Fault, the Springhill Fault is interpreted to be a south-dipping compressional structure, given that the older Ridge Brook Formation on the southeast is juxtaposed against younger strata of the Windsor, Mabou, and Pictou groups on the northwest (Fig. 4, 5, 6).

The Cosman Settlement Fault can be traced from Mount Hebron in the south to Springhill in the north, where it links with the Springhill Fault (Fig. 4). It was first recognized as a significant break by St. Peter (1989) on the southwest side of Jordan Mountain, where beds of the Memramcook Formation immediately west of the fault trace strike obliquely toward the adjacent basement. This relationship suggests that the basement and its cover of Ridge Brook Formation may have moved up relative to the Memramcook Formation, which, if true, would support the interpreted up-on-the-southeast displacement on the Springhill Fault.

North of Cosman Settlement, the Cosman Settlement Fault joins the subparallel-striking Berwick Fault (Fig. 4). These faults are visible in roadcuts along Highway 2, where conglomerates and sandstones of the Ridge Brook Formation on the east side of the fault are thrown against conglomerates and sandstones of the Bloomfield Formation on the west. Although the nearest roadcuts are several hundred metres from the interpreted fault trace location, the magnitude of displacement on the fault can be inferred from the unusually steep dips of the formations. Here, the normally shallowly dipping Ridge Brook beds dip 40°SE to 50°SE, whereas the Bloomfield beds dip 40°NW to 60°NW. In roadcuts, the Ridge Brook Formation contains minor faults that dip steeply southeast and have slickensides suggesting dip-slip displacement (Fig. 42). Conversely, minor faults in the beds of the Bloomfield Formation dip steeply northwest and have slickensides that pitch shallowly northeast, suggesting lateral oblique movement.

Berry Mills and Kennebecasis Faults

The Berry Mills Fault defines the southeastern boundary of the IMDZ (Fig. 2). Although Stewart (1941a, 1941b, 1941d) recognized the structure as a major break in the Carboniferous section, it was Gussow (1953) who named and described the Berry Mills Fault, citing topographic and seismic reflection evidence for its location and northwest dip.

Figure 42. Minor, steeply southeast-dipping faults in rocks of the Ridge Brook Formation are interpreted to be related to the Cosman Settlement and Springhill fault system. Outcrop is situated along Highway 2 (Trans-Canada Highway) northeast of Cosman Settlement.

Seismic reflection profiles from Irving/Chevron surveys in the early 1980s also indicate a steep, northwest dip.

The segment of the Berry Mills Fault between Sussex and Petitcodiac (Fig. 4, 5) was called the Petitcodiac Fault by Gussow (1953). He interpreted strata of the Cumberland and Pictou groups to overstep, and not be offset by, his Petitcodiac Fault. However, circumstantial evidence suggests that this segment of the Berry Mills Fault does cut the Pennsylvanian strata. For instance, an abrupt and nearly straight topographic break defines the northwestern limit of the Boss Point and Salisbury formations from Mount Pisgah to Petitcodiac. This post-Pictou displacement is confirmed northeast of Irishtown along Shediac River and McQuade Brook (Fig. 6), where Westphalian D and Stephanian beds of the regionally subhorizontally disposed Richibucto Formation dip 70° to 90° at the fault trace (St. Peter and Park 2006b).

The Kennebecasis Fault was named and described by Gussow (1953) but had been mapped earlier by Alcock and MacKenzie (1946), who recognized it as a major tectonic break that carried crystalline rocks of the Kingston Uplift (Fig. 2) southeastward over Carboniferous footwall rocks (Fig. 43). Unlike Gussow, Alcock and MacKenzie (1946) tentatively interpreted that the northeastern extension of their Kennebecasis Fault assumed a more northerly trend toward Jordan Mountain. Whether the Kennebecasis Fault actually links with the Berry Mills Fault (as shown on Fig. 4) is not yet known for certain.

Figure 43. View looking northwest near Norton, showing the Kennebecasis Fault trace, marked as a dotted line. The crystalline basement forms the hanging-wall highlands in the background against the Mabou Group, which underlies the field and woodland in the foreground.

Further study is needed to ascertain the possible location of the Kennebecasis Fault from Apohaqui to Smith Creek. For example, rocks of the Windsor Group immediately north of Apohaqui and Mount Pisgah on the footwall side of the fault appear to be cut off, yet the Windsor strata south of McGregor Brook are folded but otherwise little displaced. A significant post-Mabou (Late Mississippian) displacement exists on the Kennebecasis Fault, as is demonstrated southwest of Apohaqui (Alcock and Mackenzie 1946). In addition, a later, post-Pictou displacement on the Berry Mills Fault is apparent northeast of Mount Pisgah. However, these displacements are not represented by a correlative, substantial offset of the Windsor Group south of McGregor Brook.

Two possible explanations are offered for this absence: either 1) the southwestern part of the Berry Mills Fault may have been erroneously located, such that the fault actually swings westerly from Smith Creek and joins the Jordan Mountain Fault near Lower Millstream, thereby remaining on the north side of the Windsor rocks at McGregor Brook, or 2) the post-Mabou strain on the Kennebecasis Fault northeast of Apohaqui was transferred to the Penobsquis Fault (Fig. 4).

The Berry Mills Fault and apparently the Kennebecasis Fault were not active structures that gave rise to escarpments and influenced sedimentation during the Sussex–Pictou depositional cycles. Formations within the Sussex, Windsor, Mabou, Cumberland, and Pictou groups on either side of these faults show no known significant differences in lithofacies or thickness.

An unresolved issue exists regarding correlation of the Horton Group across the Kennebecasis–Berry Mills fault system. It particularly concerns trying to relate the Albert Formation in the Apohaqui–Wards Creek area of the Moncton Subbasin with the Albert Formation in the Lower Millstream–Cosman Settlement area of the Cocagne Subbasin. As described earlier under *Albert Formation*, the formation differs substantially in the two areas. In the Apohaqui–Wards Creek area it constitutes a complex lacustrine and fluvial depositional milieu, yet in the Lower Millstream–Cosman Settlement area it comprises a much simpler sequence of stacked, fining-upward fluvial deposits. It is not known whether the Albert sequences in these two areas represent deposits in two different depocentres during the Early Mississippian, but, as discussed under *Jordan Mountain Fault*, a highland area (including the Jordan Mountain exhumed basement) may have existed between the two areas during the Horton depositional cycle.

Penobsquis Fault

The Penobsquis Fault is an east- to northeast-trending break that extends from Apohaqui to Springdale and beyond (Fig. 4). Its existence and northwest dip in the Springdale area was first noted by Gussow (1953) and later observed in the Apohaqui–Penobsquis area by St. Peter (2002). Wilson et al. (2006) recently described the fault as a southeast-verging reverse structure associated with a décollement in a northeast-trending 'wall' of Windsor evaporites that occurs in the subsurface between McCully and Springdale (Fig. 4). The fault rises out of the Windsor evaporites and displaces beds of both the Mabou and Cumberland groups in surface outcrops. The Penobsquis Fault is interpreted as a splay of the southeast-verging Kennebecasis Fault.

CONCLUSIONS

Over the past few years, work in the Maritimes Basin of southeastern New Brunswick has resulted in significant advancements in palynostratigraphy that have, in conjunction with detailed field mapping, resulted in revisions to Upper Devonian–Carboniferous lithostratigraphy in the region. These refinements in biostratigraphy and lithostratigraphy have advanced, and been accompanied by, a greater understanding of major tectonic elements that controlled the location and evolution of the larger scale components (subbasins and uplifts) that comprise the Maritimes Basin complex.

Several important new findings have emerged from the foregoing work, particularly those concerning the stratigraphy and structural relationships of the Cocagne Subbasin with the Indian Mountain Deformed Zone (IMDZ) and Moncton Subbasin. The most important findings, together with their interpretations, are summarized below under stratigraphic revisions, lithofacies distributions and thicknesses, and structural observations and interpretations.

Stratigraphic Revisions

1. Fine-grained grey and red clastic strata previously assigned to the upper part of the Memramcook Formation or lower part of the Albert Formation in the IMDZ, and in the Cocagne and Moncton subbasins, have been palynologically dated as early Tournaisian and given the new name McQuade Brook Formation.

2. Greyish green to reddish brown sandstones and mudstones in the northwestern domain of the IMDZ that conformably overlie the Albert Formation bear Tournaisian Zone 2 to 3A spore assemblages and have been given the new name Indian Mountain Formation.

3. Pinkish grey, feldspar-rich conglomerates, sandstones, and red mudstones in the southwestern parts of the Cocagne and Moncton subbasins that concordantly overlie the Albert Formation also contain Tournaisian Zone 2 or 3A spores and have been assigned the name Bloomfield Formation.

4. The Horton Group and older (basement) rocks are unconformably overlain by the Sussex Group of late Tournaisian age.

5. The base of the Sussex Group in the southeastern domain of the IMDZ is marked by a coarse breccia and conglomerate of the newly defined Stilesville Formation.

6. The Stilesville Formation is succeeded by the Gautreau Formation, which has been removed from its historical inclusion in the upper part of the Albert Formation, redefined, and placed in the Sussex Group on the basis of its palynological content and lithostratigraphic position.

7. The Gautreau Formation is succeeded in the southeastern domain of the IMDZ by red mudstones of the Weldon Formation, which in turn are gradationally overlain by conglomerates and lithic sandstones of the late Tournaisian, newly named Briggs Cross Formation.

8. Conglomerates and sandstones formerly placed in the Hillsborough Formation in the Cocagne Subbasin are assigned the new name of Ridge Brook Formation and are placed in the Sussex Group on the basis of their lithologic distinctiveness, late Tournaisian palynology, and contact relationship with the overlying Windsor Group.

9. The conglomeratic Hillsborough Formation unconformably overlies the Sussex Group and concordantly underlies limestones of the Gays River and Macumber formations in the Moncton Subbasin, and accordingly is reassigned to the base of the Windsor Group, as originally mapped by Wright (1922).

10. Conglomerates and sandstones that flank the New Brunswick Platform from Belleisle Creek to Springhill and that previously were mapped as the base of the Horton Group (Memramcook Formation) are assigned to the undivided Mabou Group. This reassignment is based on the observation that they lie above limestone of the Windsor Group in the Springhill area and are laterally continuous with the Shin Formation (Mabou Group) on the New Brunswick Platform.

Lithofacies Distributions and Thicknesses

1. The Memramcook Formation at the base of the Horton Group appears in the subsurface in the northeastern Cocagne Subbasin as a thick, coarse, polymictic conglomerate unit adjacent to the Belleisle Fault, suggesting proximity to the northwestern margin of the Cocagne Subbasin.

2. The fluvial sequence of the Albert Formation in the southwestern Cocagne Subbasin contrasts with the predominantly lacustrine sequence of the same formation in the Moncton Subbasin.

3. The thick sequences of the Albert and Indian Mountain formations (Horton Group) that lie adjacent to the North River Fault in the northwestern domain of the IMDZ are not basin-margin sequences and therefore do not represent the southeastern depositional limit of a Horton basin.

4. The lithofacies distributions, thicknesses, and transport directions of the coarse facies (Stilesville, Briggs Cross, Ridge Brook, and Mill Brook formations) and fine facies (Gautreau, Weldon, and Dutch Valley formations) of the Sussex Group indicate that the coarse facies circumscribe the fine facies. This configuration suggests that these coarse and fine units represent the proximal and distal facies, respectively, within a single basin centred in the Petitcodiac area during the Sussex depositional cycle.

5. The initiation of Windsor deposition is reflected by terrestrial clastics of the Hillsborough Formation, and by the lithofacies distribution of the overlying Windsor carbonates and evaporates. Both features indicate that the axis of the Windsor depocentre lay between the Westmorland Uplift and the Caledonia Uplift and extended from the Sackville Subbasin to the Sussex–Lower Millstream area (see McCutcheon 1981). Accordingly, the location of the Windsor depocentre in southern New Brunswick appears to have been controlled more by the location of contemporaneously exposed crystalline highlands than by the location of the older Horton or Sussex basins, or by the location of major faults that partially preserved those basins.

6. The differing thicknesses and distributions of the coarse and fine lithofacies of the Mabou Group were studied only summarily. However, as with the lithofacies configuration of the Sussex Group, these contrasting features imply that a single Mabou depocentre was centred near Petitcodiac, and that its margins were controlled mainly by the New Brunswick Platform, and the Westmorland and Caledonia uplifts.

Structural Observations and Interpretations

Several important new structural observations, interpretations, and conclusions have arisen from the foregoing stratigraphic and lithofacies findings. These observations and interpretations have a particular bearing on defining relationships between the Cocagne and Moncton subbasins, the IMDZ, and their bounding uplifts in the Maritimes Basin of New Brunswick.

1. The Cocagne Subbasin is a structural basin and is characterized primarily by a thick sequence of distinctive strata of the Horton Group and to a lesser degree by sequences of the Sussex and Windsor groups that contrast with sequences of the Sussex and Windsor groups in the IMDZ and Moncton Subbasin.

2. The Belleisle Fault was a long-persistent structure that delimits the southern boundary of the New Brunswick Platform. It had a controlling influence on the northwestern margin of the depocentres that accumulated clastic sediment during the Horton, Sussex, and Mabou depositional cycles.

3. The IMDZ is a northeast-trending, uplifted inversion zone (block) bounded by the Smith Creek and Berry Mills faults, and comprising two contrasting structural domains separated by the North River Fault. The northwestern domain exposes a thick sequence of the Horton Group, while the southeastern domain is floored by basement rocks that are overlain by thin outliers of preserved Horton rocks and an unconformably overlying thick sequence of the Sussex Group.

4. In the southwestern part of the report area, crystalline rocks on the northwest (hanging-wall) side of the Jordan Mountain Fault may reflect the existence of a larger, buried basement block that was uplifted along the Kennebecasis–Berry Mills fault system. The basement block may have acted as a highland during the Horton depositional cycle, which would explain the contrast in lithofacies of the Albert Formation in the southwestern Cocagne Subbasin and Moncton Subbasin.

5. The section of Horton strata on the northwest side of the North River Fault defines the northwestern domain of the IMDZ and has an unclear relationship with the Cocagne and Moncton subbasins as herein defined. Spatially, it lies on the northwestern side of the

Westmorland Uplift (Fig. 2) and so is more logically related to the Cocagne Subbasin, which lies buried beneath strata of the Pictou and Sussex groups and immediately north of the Smith Creek Fault. However, lacustrine strata of the Albert Formation (Horton Group) in the IMDZ are unlike Albert beds in the Lower Millstream area or buried Albert beds in borehole G-56 in the Monteagle area, both in the Cocagne Subbasin. Instead, the Albert sequence in the IMDZ is more comparable to the lacustrine Albert sequence in the Moncton Subbasin (St. Peter 1989, 1992).

6. The southeastern domain of the IMDZ in The Gorge area is interpreted as a raised block of the Westmorland Uplift because, like that uplift, it is characterized by the absence or paucity of Horton strata unconformably beneath sequences of younger Carboniferous rocks.

7. At Jordan Mountain, the basement covered by strata of the Sussex Group was inverted on southeast-verging faults. In contrast, at The Gorge, the basement covered by strata of the Sussex Group was inverted by mainly northwest-verging faults.

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